

ENVIRONMENTAL AND PEDOGENIC CHANGE
IN THE CENTRAL GREAT PLAINS
FROM THE MIDDLE WISCONSINAN TO THE PRESENT

By

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B.A., University of Kansas, 1997

Submitted to the graduate degree program in Geography and the
Faculty of the Graduate School of the University of Kansas
In partial fulfillment of the requirements for the degree of
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For my husband Scott,
who wanted this more than anyone

Acknowledgements

This research could not have been completed without the support of many people. I would like to thank my co-authors Bill Johnson, Joe Mason, Dave May, and Gwen Macpherson for their substantial contributions. Technical and laboratory assistance was provided by the staff of the KU Soil Geomorphology Laboratory: Josh Campbell, Casey McLaughlin, Aura Whitcomb, Stephanie Hewett, and Julie Robinson. Additionally, Roxane Fagan, Matt Hohmann, and the staff of the Konza Tallgrass Prairie LTER site provided analytical and logistical help. Years of field work employed the efforts of many people without whom I would still be trudging in the remote reaches of the central Great Plains: Scott Eudaly, Glenn Willey, Bill Kastner, Fred Magley, Tobe Zwegardt, Larry Gatto, Gabby Blair, Jason Blair, and Ben Johnson, Kyeong Park, Doug Kolom, and of course my advisor, Bill Johnson. Thank you to Dave May, Glen Fredlund, David Grimley, my dissertation committee, and several anonymous reviewers whose editorial comments improved this work greatly. The willingness of the many landowners and county officials to allow property access was critical to the research, as was the hospitality of the people of St. Francis, Kansas.

Funding for this dissertation was provided by a National Science Foundation Doctoral Dissertation Improvement Grant (BCS-0118012), a United States Geological Survey Educational Mapping Program (EDMAP) grant, and a Geological Society of America Robert K. Fahnestock Memorial Research Award granted to the author. Additional support was provided by the United States Army Corps of Engineers ERDC BT-25 Program through Bill Johnson and Gwen Macpherson, and by a National Science Foundation grant (SBR-9601170) and the General Research Fund, University of Kansas awarded to Bill Johnson. John Blair funded stable isotope analyses on Konza samples.

I owe great thanks to my husband, Scott Eudaly, my daughter, Katherine, my extended family, and my church congregation for their endless support of my doctoral efforts. Thank you to my advisor, Bill Johnson, for the investment of over a decade of his time and energy, his tireless field work, his boundless patience, and his sharp wit. He taught me to “see” the landscape through the windshield of a Ram Charger.

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ABSTRACT

During the middle Wisconsinan, the Gilman Canyon Formation (GCF), consisting of three loess units and three soils, formed on the loess plateaus of the central Great Plains about 40-25 ka. Stable carbon isotope analysis of the lower two GCF loess units (L1 and L2) at the type locality in southwestern Nebraska, Buzzard's Roost, revealed a mixed C₃/C₄ grassland. Strongest pedogenesis (GCF S3) and C₄-dominance correlated strongly with a peak in summer insolation for the northern latitudes. Soil 2 (S2), L3 and a weakly-expressed S1 developed successively as insolation values declined and $\delta^{13}\text{C}$ values synchronously dropped back, reflecting the emerging boreal environments of the Late Wisconsinan and deposition of the Peoria Loess.

As the Pleistocene ended, Peoria Loess deposition waned sufficiently for pedogenesis to prevail across the central Great Plains, resulting in development of the upland Brady Soil and temporal equivalents in other landscape positions. Climate warmed rapidly between early (13-11 ka) and late (10-9 ka) development of the Brady Soil. Early Holocene temperatures, inferred from carbon isotopic signatures of the Brady Soil, rose 8-10°C above those of the Late Pleistocene.

The Holocene record in loess sections of the central Great Plains generally lacks resolution for environmental reconstruction, but canyons systems of the upper Republican River suggest massive Altithermal erosion from *c* 8-5 ka. T-2 alluviation,

with periodic soil formation, ensued from *c* 5 ka to onset of the Medieval Warm Period (*c* 1 ka), when erosion initiated a fresh gully cycle.

Modern soils at Konza in northeastern Kansas suggest that $\delta^{13}\text{C}$ in C_4 environments such as those of the GCF S3 and the Brady Soil are insensitive to minor variability with landscape position and can therefore be confidently used for paleoenvironmental reconstruction despite the often unknown paleolandscape position of sampled paleosols. Further research is needed, however, to resolve the issue of low surface soil $\delta^{13}\text{C}$ values relative to biomass found in modern soils, a phenomenon that could lead to an under-representation of C_4 plant contributions.

CHAPTER 1: INTRODUCTION

Global environmental changes of the last few to several 100 kyr are well-represented in accumulated layers of glacial ice and benthic marine sediments, and in particular provide a detailed record of the last 100 kyr (Figure 1). Proxy data (*e.g.* $\delta^{18}\text{O}$) from these archives are frequently correlated with data derived from their continental cognates. Elevated ocean temperatures of the Last Interglacial—the Sangamon (Marine Isotope Stage 5: MIS 5)—were indicated by $\delta^{18}\text{O}$ values similar to those of the present, but exhibited a downward trend during the Wisconsinan Glaciation (MIS 4-2), reaching a minimum during the Last Glacial Maximum (LGM) (Figure 1b). Following the Last Glacial Maximum, temperatures warmed abruptly, increasing oceanic-core ^{18}O concentrations. Holocene (MIS 1) temperatures remained relatively high and exhibited only minor fluctuations. Ice core records from GISP2 express greater variability by recording the $\delta^{18}\text{O}$ signal of yearly precipitation (Figure 1a). Abrupt warming and gradual cooling cycles (Dansgaard-Oeschger cycles) occurred throughout MIS 5-2, but are particularly prominent features of MIS 3 (Bond *et al.*, 1993). Each cycle ended with the release of great quantities of sea ice into the North Atlantic, *i.e.* Heinrich Events (Heinrich, 1988). Heinrich events appear to have continental expressions (*e.g.* Grimm *et al.*, 2006; Wang *et al.*, 2001). The Younger Dryas event (Alley *et al.*, 1993), a period of abrupt cooling and warming at the MIS 2/MIS1 boundary, is well-manifested in the GISP2 core data, but the averaging of multiple benthic core records (Figure 1b) has blurred this event (Figure 1a).

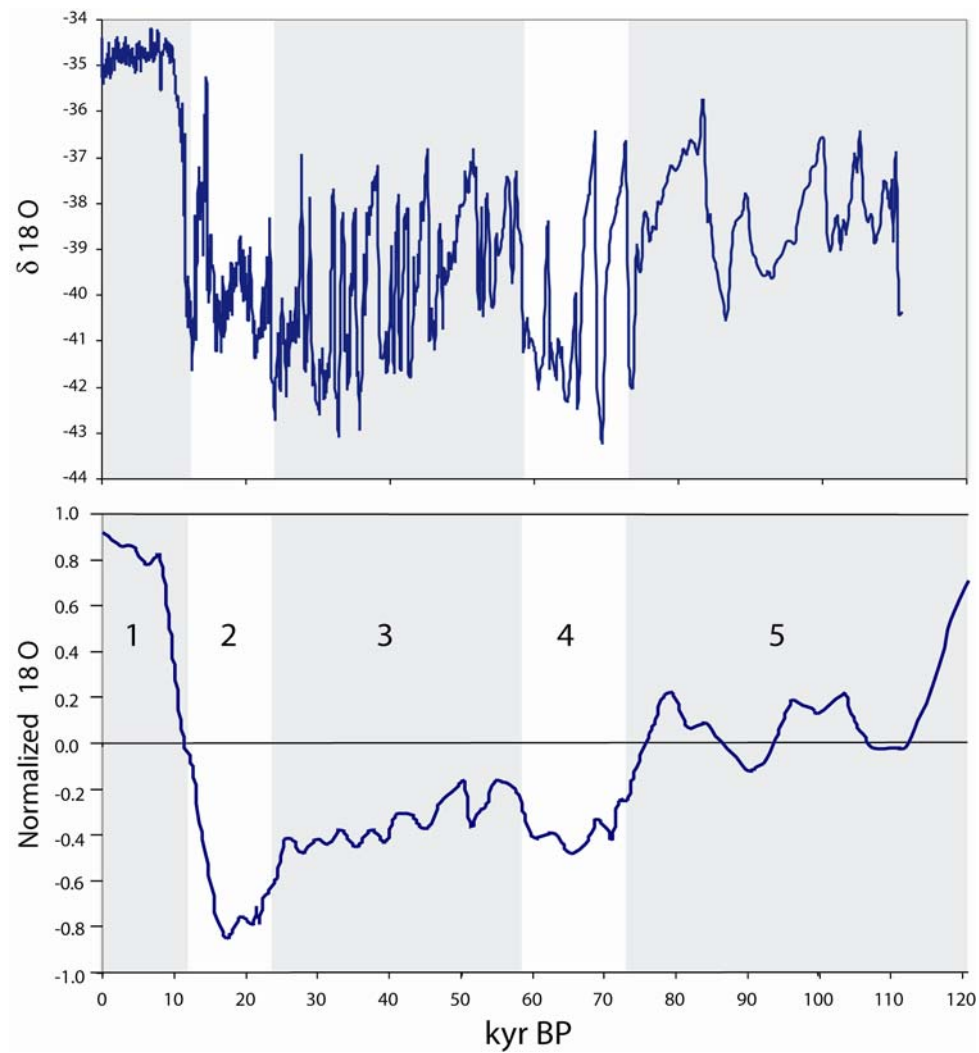


Figure 1. Marine isotope stages of the Sangamon Interstade (MIS 5), Wisconsinan Stadial (MIS 4-2) and Holocene (MIS 1) as identified with $\delta^{18}\text{O}$ of the GISP2 core (Stuiver and Grootes, 2000) and a normalized compilation of North Atlantic benthic sediment cores (Martinson *et al.*, 1987).

In the Great Plains, accumulated loess deposits provide a nearly-continuous record of environmental change, which serves as a terrestrial archive of paleoclimatic data. Loess and intercalated soils provide information via a wide array of proxies, *e.g.*, rock magnetism, stable isotope ratios, biogenic opal, charcoal, elemental composition, granulometry, and faunal remains.

Geographically, this research was focused on loess and loess-derived soils in the central Great Plains (Kansas and Nebraska). Further, though Loveland and Pre-Loveland loess units are expressed in the thick bluffs along the Platte River in Nebraska and elsewhere, this research explored the record from the Middle Wisconsinan to the present. The Gilman Canyon Formation (GCF) is a well-recognized loess unit within Nebraska and Kansas featuring a prominent soil complex associated with MIS 3, an interstade within the Wisconsinan Glaciation. This document introduces detailed stratigraphy and dating for the GCF at the type locality, Buzzard's Roost, southwestern Nebraska. Stable carbon isotope ratio analysis, room-temperature rock magnetic parameters, and chemical and physical soil parameters were employed to provide evidence of environmental change throughout GCF deposition and pedogenesis.

Sites across Kansas and Nebraska were used to characterize the environment of the Pleistocene-Holocene transition as captured in the loess-derived Brady Soil. Approximately concurrent with the Younger Dryas event, genesis of this widespread paleosol occurred along with rapid climate warming and landscape stability. Paleowind flow out of the northwest in the Late Wisconsinan accounted for the thick

accumulations of Peoria Loess along the Platte River in Nebraska and decreasing thicknesses to the south and east (Muhs and Bettis, 2000). With the retreat of the Laurentide ice sheet at the onset of the Holocene, wind direction changed to favor southerly winds. The Brady Soil formed during this transition and records climatic warming, decreased effective moisture for plants, and increasing seasonality in the central Great Plains. Initiation and termination of Brady Soil development was examined with respect to latitude and longitude to evaluate possible time-transgression. Rock magnetic parameters (susceptibility and frequency dependence of susceptibility) were used to characterize weathering profiles of several Brady Soil sites and provide insight on longitudinal trends of paleosol pedogenesis. Vegetation change through the Pleistocene-Holocene transition was reconstructed for three sites along an east-west transect using stable carbon isotope ratio data.

Though Holocene temperature variations have been minor with respect to the last 100 kyr, fluctuations have produced major geomorphic responses in fluvial systems and sand sheets in the central Great Plains. Holocene environmental changes, difficult to derive from the decreased loess influx alone, were identified through the cut and fill cycles of canyons dissecting the loess plateau along the upper Republican River of Kansas and Nebraska. Hay Canyon was used to generate a model of sediment movement and storage, and ages of buried soils in nearby canyons corroborated the model. The model, which correlates with other regional records of environmental change, highlights several periods of soil development and widespread erosion 3-5 ka.

The last study reported within this dissertation compares above-ground biomass associated with modern soils to the proxy signal contained therein. Specifically, a key paleovegetational proxy method used throughout this study, that of stable carbon isotope analysis, was assessed for its reliability, *i.e.*, do stable isotope values derived from paleosols accurately render the record of the prevailing vegetation. A modern analog study was conducted on the Konza Tallgrass Prairie Long Term Ecological Research (LTER) site in east-central Kansas. Vegetation and soil were sampled from a C₄ plant-dominated prairie with similar $\delta^{13}\text{C}$ values to Gilman Canyon Formation soil and the Brady soil. These two paleosols from the Beisel-Steinle site in central Kansas were compared to modern findings in order to interpret past climatic changes. Previous carbon isotopic studies have examined regional variation, but the effect of microclimate on the $\delta^{13}\text{C}$ values of above-ground biomass and corresponding surface soils has not received similar attention.

CHAPTER 2: Stratigraphy and Environmental Reconstruction at the Middle Wisconsin Gilman Canyon Formation Type Locality, Buzzard's Roost, Southwestern Nebraska

(published as Johnson *et al.*, 2007a)

Abstract

The Middle Wisconsin Gilman Canyon Formation at the Buzzard's Roost type locality in southwestern Nebraska was investigated to document the stratigraphy and to reconstruct the environmental and climate record. The Gilman Canyon Formation includes three loess units and three soils, with radiocarbon ages constraining its age between about 40 ka and 25 ka. Stable carbon isotope ratios, magnetic susceptibility, and carbon content were used to define and characterize soils within both the Gilman Canyon Formation and underlying Illinoian Loveland Loess. At the height of soil development within the Gilman Canyon Formation, climate was supporting a C₄-dominated grassland, with July temperatures equal to or exceeding those of today. Soil-forming intervals within the Loveland Loess, including the Sangamon Soil, also exhibited increases in C₄ biomass relative to minimally-pedogenic loess. Climate, as recorded in the Gilman Canyon Formation, is corroborated by regional proxy data. The formation accumulated during Marine Isotope Stage (MIS) 3, and concurrent soil formation coincided with a summer insolation maximum.

Introduction

Loess research in the central Great Plains (Nebraska, Kansas, and eastern Colorado) has undergone a renaissance in recent decades, resuming work begun largely during the middle to late 1940s. Notable early regional studies of loess stratigraphy include those by Schultz and Stout (1945, 1948) in Nebraska and by Frye *et al.* (1948) and Frye and Leonard (1952) in Kansas. Recent research has been driven to a large extent by the realization that the loess record can be a continental parallel of the deep sea and ice core records.

Contemporary research in the central Great Plains frequently makes reference to type localities established from the 1940s to the 1960s, and, given the recognition of these sites, current analytical research tools should be focused on these type sections to better characterize stratigraphic relationships, chronology and environmental history. Some classic type sections have been re-investigated in recent years, such as the Illinoian Loveland Loess at the Loveland paratype locality in eastern Iowa (Muhs and Bettis, 2000; Forman and Pierson, 2002; Bettis *et al.*, 2003a, 2003b) and the Brady Soil and Holocene Bignell Loess at Bignell Hill, southwestern Nebraska (Maat and Johnson, 1996; Muhs *et al.*, 1999a; Bettis *et al.*, 2003a, b). Presented here are new radiocarbon, isotope, magnetic and sedimentological information for the Gilman Canyon Formation (GCF) type locality at Buzzard's Roost, southwestern Nebraska. The GCF consists of a loess unit corresponding with most of marine isotope stage 3 (MIS 3) which is a time interval that has received little attention in the Great Plains.

Loess Stratigraphy in the central Great Plains

Late Quaternary loess of the central Great Plains is comprised of five units including the basal pre-Illinoian loesses, Loveland Loess (Illinoian), GCF (Middle Wisconsinan), Peoria Loess (Late Wisconsinan), and Bignell Loess (Holocene) (Figure 2). Each of these units is recognized as a formation in Nebraska (Reed and Dreeszen, 1965) and has been correlated with the loess stratigraphy of Kansas (Frye and Leonard, 1951, 1952). Intercalated soils occur throughout the loess sequence, providing indications of periodic landscape stability and insight into paleoenvironmental conditions. Pre-Illinoian loess is recognized by the existence of one or more interstadial soils (*e.g.*, “Yarmouth Geosol”: Wilman and Frye, 1970) and of underlying Quaternary volcanic ash at some sites, such as at the Eustis ash pit (Schultz and Stout, 1945; Fredlund *et al.*, 1985) and Buzzard’s Roost localities (Reed and Dreeszen, 1965). At Eustis, ash underlying pre-Illinoian loess was identified as the Lava Creek B (Pearlette Type O), dated to 620 ka (Izett and Wilcox, 1982). At this same location, a carbonate nodule from the Btk horizon of a well-developed paleosol immediately below the Loveland Loess provided a U-series age of 184 ka (Muhs *et al.*, 1999b).

Loveland Loess has been documented throughout the Missouri, Mississippi, and Ohio River basins (Wilman and Frye, 1970; Ruhe and Olson, 1980). Maat and Johnson (1996) reported TL ages averaging 163 ka on the upper Loveland Loess at Eustis, which are in agreement with ages obtained by Forman *et al.* (1992) from the Loveland Loess paratype locality in western Iowa and by Oviatt *et al.* (1988) on

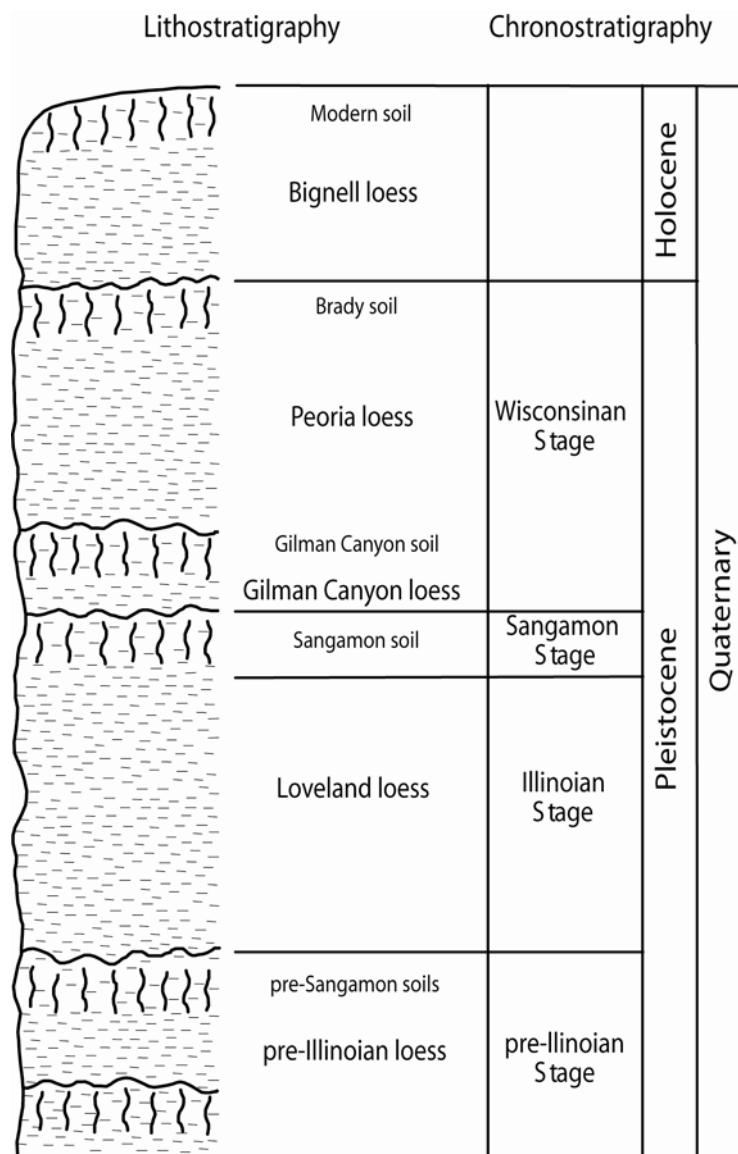


Figure 2. Composite Late Quaternary loess stratigraphy of the central Great Plains. The Sangamon Soil is about 120 kyr old; the Gilman Canyon Soil is about 40-25 kyr old; and the Brady Soil is about 13-9 kyr old. Vertical scaling varies.

Loveland Loess in east-central Kansas. The most remarkable feature of the Loveland Loess is the Sangamon Soil, formed within the upper 1-2 m during MIS 5: the last interglacial (Markewich *et al.*, 1998). This paleosol typically features a structural and argillic B horizon, carbonate-rich B or C horizons, and yellow-red hues.

The GCF is a relatively thin (<4 m), leached loess unit dominated by dark, organic-rich A horizons indicating two or more episodes of cumulic soil development. Regional expression indicates this GCF pedocomplex, like the Sangamon Soil, is a geosol or composite geosol, which appears coeval with the Farmdale Geosol in Illinois (Follmer, 1983). It rests unconformably on the Sangamon Soil, and, at localities where the loess is thinnest, pedogenesis has welded it to the underlying Sangamon Soil. Transition to the overlying Peoria Loess is often difficult to discern visually, but color brightens gradually due to a drop in soil organic carbon (SOC) content, structure becomes more massive, and carbonate content increases. Radiocarbon and TL ages reported for the GCF indicate that its age range is about 40-22 ka (Johnson, 1993; Martin, 1993; May and Holen, 1993; Pye *et al.*, 1995; Maat and Johnson, 1996; Muhs *et al.*, 1999b; Muhs *et al.*, 2008).

The overlying late Wisconsinan Peoria Loess is the most widely distributed of all loess units and thickness ranges up to 50 m in southwestern Nebraska (Maat and Johnson, 1996; Roberts *et al.*, 2003). It is typically calcareous and massive, with weakly developed buried soils. Radiocarbon ages obtained from the Peoria Loess range from about 21,000 to 10,000 ^{14}C yr BP (Wells and Stewart, 1987; Johnson, 1993; Martin, 1993; May and Holen, 1993; Feng *et al.*, 1994a, b; Maat and Johnson,

1996), with TL and OSL ages providing a range of about 25-12 ka (Pye *et al.*, 1995; Maat and Johnson, 1996; Roberts *et al.*, 2003).

Capping the Peoria Loess is the well-developed Brady Soil, which is only recognizable where the overlying Holocene Bignell Loess is present. Otherwise, the Brady Soil has either been eroded prior to modern soil development or been incorporated into the modern soil profile. Radiocarbon ages from the Brady Soil range from about 12,000 to 8,000 ^{14}C yr BP (Johnson, 1993; Martin, 1993; Maat and Johnson, 1996; Muhs *et al.*, 1999a; Johnson and Willey, 2000; May and Holen, 2003). TL and OSL ages from the Bignell Loess provide a spectrum of Holocene ages, from about 10 ka to 0.7 ka (Pye *et al.*, 1995; Maat and Johnson, 1996; Mason and Kuzila, 2000; Mason *et al.*, 2003; Miao *et al.*, 2005).

Gilman Canyon Formation

Previous studies

What is now defined as the GCF was first recognized as part of the “Citellus zone” (Schultz, 1934), a stratigraphic interval dominated by burrows and skeletal remains of rodent taxa, particularly the ground squirrel *Spermophilus richardsonii* (*Spermophilus* has since replaced *Citellus* as genus name). Schultz and Stout (1945), Condra *et al.* (1947) and Frye and Leonard (1951) noted burrows extending from the top of the present GCF into the uppermost Loveland Formation. Recently, Tobin (2004a, b) reported that burrows originate within the soil zone of the GCF and often extend through the Sangamon Soil into the Loveland Loess. The GCF, with its

dominant pedogenesis, was originally believed to be the A horizon of the Illinoian Sangamon Soil (Condra *et al.*, 1947; Swineford and Frye, 1951; Thorp *et al.*, 1951), and, in Nebraska, the soil couplet was designated the Loveland Soil by Condra *et al.* (1947). Similarly, Thorp *et al.* (1951) recognized an unnamed Early Wisconsinan loess above the Sangamon Soil at Yankee Hill Brickyard in southeastern Nebraska. Subsequently, Reed and Dreeszen (1965, p.40) introduced the name Gilman Canyon Formation for the “comparatively thick zone of humic silt” above the Sangamon Soil and designated Buzzard’s Roost as the type locality. Although no radiocarbon ages were available, they were convinced by the stratigraphic expression that the GCF was not the A horizon of the Sangamon Soil, but rather a discrete Early Wisconsinan deposit.

Buzzard’s Roost is located in the West Fork of Gilman Canyon, one of several such features in the loess region south of the Platte River valley in southwestern Nebraska (Figure 3). The section, as originally described, is a composite constructed largely from overlapping profiles along the cut of a steep dirt road connecting the West Fork bottoms with the loess uplands. Schultz and Stout (1945) first recorded the Buzzard’s Roost locality, which was subsequently documented by Frankel (1956), Schultz and Tanner (1957), Reed and Dreeszen (1965), Schultz (1968) and Schultz and Martin (1970).

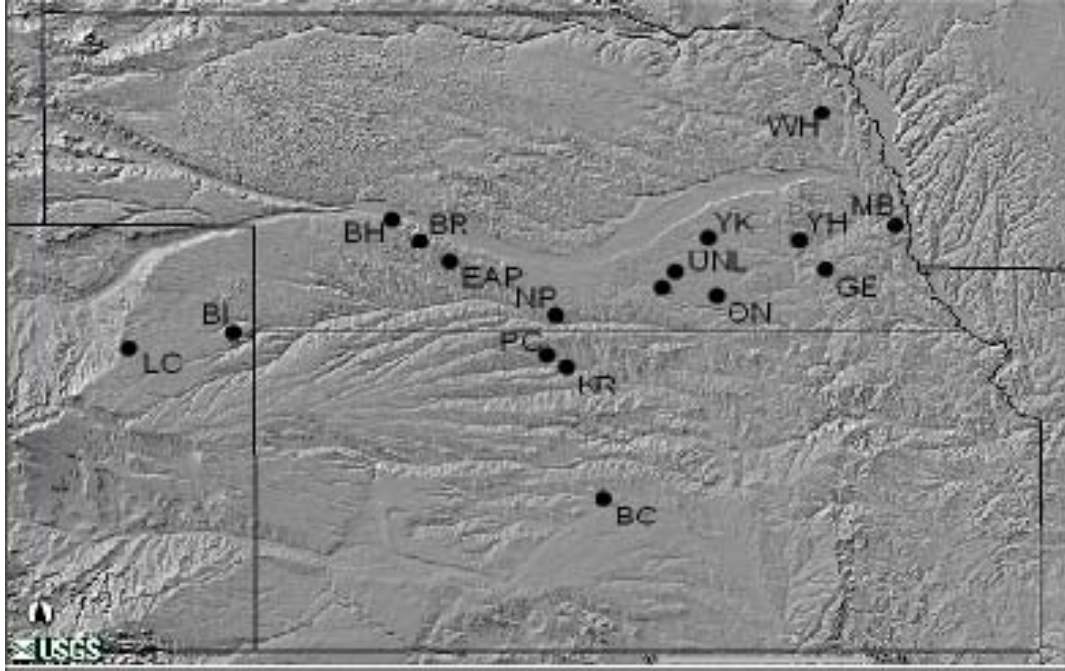


Figure 3. Shaded relief digital elevation model of the central Great Plains, loess distribution, and Gilman Canyon Formation localities referenced in Table 1. Sites include Barton County Landfill (BC), Beecher Island (BI), Bignell Hill (BH), Buzzard's Roost (BR), Eustis Ash Pit (EP), General Excavating (GE), Kirwin Railroad cut (KR), Last Chance (LC), Maas Drive (MD), Naponee (NA), Ong (ON), Phillips County Landfill (PC), Spring Ranch (SR), UNL South-Central Agricultural Farm (UF), Winslow Hill (WH), Yankee Hill Brick Plant (YH) and York (YK). Image source: USGS

Pedologic expression

Multiple periods of soil formation have been documented within the GCF beyond the type locality. Muhs *et al.* (1999a) recognized and radiocarbon dated two distinct periods of soil formation within the GCF at Bignell Hill, Nebraska, about 33 km northwest of Buzzard's Roost. At the Last Chance site in east-central Colorado, Muhs *et al.* (1999b) documented two separate soils within a Gilman Canyon-age loess body. While Muhs *et al.* (1999b) reported only one period of soil development at the Beecher Island site in eastern Colorado, two closely-spaced, but discrete A horizons were observed concurrent with this study.

At other sites in the region where the GCF has developed on slopes, rather than on relatively stable uplands, A horizons tend to be widely separated, often by a meter or more. In south-central Nebraska, May and Souders (1988) recognized multiple, discrete A horizons distorted by slope processes at the Johnson Lake road-cut locality, whereas Johnson (1996) documented three individual A horizons on a paleoslope adjacent to a Late Wisconsinan slump feature at Harlan County Lake.

Regional chronology

Radiocarbon and TL ages from the GCF have been reported for several upland sites within the central Great Plains (Table 1). Three previously reported radiocarbon ages from Buzzard's Roost range from 21,290 ^{14}C yr BP (May and Holen, 1993) to 32,000 ^{14}C yr BP (Dreeszen, 1970), while ages from localities elsewhere in the region

Table 1. Numerical ages from upland Gilman Canyon Formation localities in Nebraska, Kansas and Colorado

Locality a	Sample ID	Source	¹⁴ C Age yr BP	Reference
Barton Co. landfill	Tx-5908	total SOC	24,360±570	Johnson, 1993
Beecher Island	CAMS-23132	humic acid b	20,520±90 c	Muhs <i>et al.</i> , 1999a
	ISGS-5624	residual	21,140±200 c	
Bignell Hill	CAMS-26401	humic acid b	30,770±210 c	Muhs <i>et al.</i> , 1999a
	CAMS-24346	humic acid b	40,600±1100 c	Muhs <i>et al.</i> , 1999a
	Tx-7706	total SOC		
	Tx-7707	total SOC		
Buzzard's Roost	CPL-14			
	Beta-26826	total SOC	21,290±290	May and Holen, 1993
	I-2188	total SOC	27,900±1100	Dreeszen, 1970
Eustis Ash Pit	I-1851	total SOC	32,000±2000	Dreeszen, 1970
	Tx-6633	total SOC	25,090±590 c	Johnson, 1990
	Tx-7356	total SOC	33,120±1020 c	Maat and Johnson, 1996
	Tx-7357	total SOC	36,000±2230 c	Maat and Johnson, 1996
	CPL-9			Maat and Johnson, 1996
General Excavation	Tx-8212	total SOC	23,460±380 c	Mandel and Bettis, 1995
	Tx-8213	total SOC	28,280±590 c	Mandel and Bettis, 1995
Kirwin RR cut	Tx-6747	total SOC	25,500±820 c	Johnson, 1993
Last Chance	CAMS-23133	humic acid b	21,060±100 c	Muhs <i>et al.</i> , 1999a
	CAMS-23134	humic acid b	22,090±100 c	Muhs <i>et al.</i> , 1999a
	CAMS-23135	humic acid b	22,940±120 c	Muhs <i>et al.</i> , 1999a
	CAMS-10190	charcoal	25,340±260	
Maas Drive	Beta-33940	total SOC	19,770±590	Mandel and Bettis, 1995
Naponee	Beta-33941	total SOC	29,870±1650	Souders and Kuzila, 1990
Ong West	Beta-24268	total SOC	22,590±280	Souders and Kuzila, 1990
	Beta-23457	total SOC	24,990±430	Kuzila and Lewis, 1993
	Beta-23456	total SOC	26,140±530	Kuzila and Lewis, 1993
Phillips Co. landfill	Tx-6746	total SOC	24,910±770 c	Johnson, 1993
Spring Ranch	Beta-20105	total SOC	21,140±220	Kuzila, 1988
	Beta-20104	total SOC	23,850±290	Kuzila, 1988
UNL Ag. Farm	Beta-20102	total SOC	20,220±330	Kitchen, 1987
Winslow Hill	I-2191	total SOC	23,000±660	Dreeszen, 1970

Table 1. continued

Locality a		Sample ID	Source	¹⁴ C Age yr BP	Reference
Yankee Hill		I-2189	total SOC	26,900±1000	Reed <i>et al.</i> , 1966
		I-2190	total SOC	34,900±2100	Reed <i>et al.</i> , 1966
York	Southwest	Beta-12273	total SOC	20,940±240	Krueger, 1986
	West	Beta-12272	total SOC	23,740±220	Krueger, 1986
	North	Beta-12274	total SOC	28,350±610	Krueger, 1986

a see Figure 3 for site locations

b for humic acid extraction technique, see Abbot and Stafford, 1996

c corrected for isotopic fractionation

range between 40,600 ^{14}C yr BP (Muhs *et al.*, 1999a) and 19,770 ^{14}C yr BP (Souders and Kuzila, 1990).

Presumed lithostratigraphic correlatives of the GCF have yielded similar ages. At the paratype locality for the Loveland Loess in Iowa, the Pisgah Formation was dated between 46 ka and 23 ka using various luminescence approaches (Forman *et al.*, 1992; Forman and Pearson, 2002; Bettis *et al.*, 2003a, 2003b). The Farmdale Geosol is developed within the upper 50-60 cm of the 4 m-thick Pisgah Formation (Bettis *et al.*, 2003a). Another Gilman Canyon correlative is the Roxana Silt (loess) of the upper Mississippi River valley and other major valleys to the east (Curry and Follmer, 1992). The Roxana Silt was generally deposited 55,000-27,000 ^{14}C yr BP (Leigh and Knox, 1993; Leigh, 1994). Subsequently, the Farmdale Geosol developed into the upper Roxana Silt from about 28,000 to 25,000 ^{14}C yr BP.

Methods

Site stratigraphy was accessed through both a profile prepared along a road cut exposing the GCF soil zone and an 18.5 m-long core from the roadbed upslope from the profile. Ten bulk radiocarbon samples from the profile exposure and three accelerator mass spectrometer (AMS) samples from the core were pretreated by disaggregating in de-ionized water, wet sieving to isolate the silt and clay fraction (<63 μm), and drying at 60 $^{\circ}$ C; carbonates were removed during pre-treatment for radiocarbon analysis.

Seven samples were collected from the profile to characterize bulk particle-size distribution from the Gilman Canyon soils to Peoria Loess. Samples were analyzed using a Coulter LS100Q laser diffraction particle-size analyzer after pretreatment with sodium acetate (adjusted to pH 5 with acetic acid) to remove carbonates (Kunze and Dixon, 1986) and with hydrogen peroxide to remove organic matter. Total, organic and inorganic (carbonate) carbon content and total nitrogen content were determined on core samples using a LECO CN combustion analyzer. In addition, organic carbon content was obtained on core samples with a Carlo Erba elemental analyzer during stable carbon isotope analysis.

Stable carbon isotope analysis

Stable carbon isotope ratios have been widely adopted in recent years to reconstruct paleoenvironments, and previous studies have applied this approach to aeolian sediments in the central Great Plains (Arbogast and Johnson, 1998; Muhs *et al.*, 1999a, b; Johnson and Willey, 2000; Feggestad *et al.*, 2005). Plants, including grasses, fix carbon from atmospheric CO₂ by one of three pathways, two of which are most relevant in the central Great Plains: the Calvin-Benson pathway (C₃) and Hatch-Slack pathway (C₄). These photosynthetic pathways allow plants differing levels of discrimination against incorporating ¹³C into tissue (Smith and Epstein, 1971). C₄ plants (warm-season grasses) produce a range in δ¹³C values of about -17‰ to -10‰, with an average of -13‰, whereas C₃ plants (cool-season grasses, most trees and shrubs) range between -32‰ and -20 ‰ and average -27‰ (Ode *et al.*, 1980;

O'Leary, 1988). Broad-scale patterns in the C₃-C₄ grass distribution are climatically forced, and several researchers have demonstrated a correlation between C₃-C₄ distribution and growing-season temperature in North America (Terri and Stowe, 1976; Fredlund and Tieszen, 1997a; Follett *et al.*, 1997a, b).

Following treatment with 1N HCl, profile and core samples were assayed using a ThermoFinnigan Delta Plus mass spectrometer. To determine the approximate percentage of SOC contributed by in-ground biomass of C₄ plants (vs. C₃ plants), various simple mixing models have been used by many (*e.g.*, Nordt *et al.*, 1994; Boutton *et al.*, 1998; Feggestad *et al.*, 2005). The equation, first derived by Ludlow *et al.* (1976) may be expressed as:

$$X = (\delta^{13}\text{C}_{\text{soc}} - \delta^{13}\text{C}_{\text{C3}}) / (\delta^{13}\text{C}_{\text{C4}} - \delta^{13}\text{C}_{\text{C3}})$$

where X is the percentage of soil organic carbon from C₄ plant material, $\delta^{13}\text{C}_{\text{soc}}$ is the derived $\delta^{13}\text{C}$ value of organic carbon in the soil; $\delta^{13}\text{C}_{\text{C3}}$ is the mean of C₃ plants (-27‰); and $\delta^{13}\text{C}_{\text{C4}}$ is the mean $\delta^{13}\text{C}$ of C₄ plants (-13‰). Resulting percentages are only approximations due to differences in biomass production, historic changes in $\delta^{13}\text{C}$ of atmospheric CO₂, differential preservation, microbial respiration, and other factors (Nordt *et al.*, 1994; Boutton, 1996).

Magnetic susceptibility

Untreated samples were loaded into 8 cm³ plastic cubes for room-temperature measurement of magnetic susceptibility (χ). Mass-specific low- (0.47 kHz) and high- (4.7 kHz) frequency magnetic susceptibility measurements were collected with a Bartington susceptibility system and are expressed in 10⁻⁸m³/kg. Low-frequency susceptibility (χ) provides an indication of the bulk ferrimagnetic mineral composition and, to a lesser extent, the ferrimagnetic mineral grain size. Frequency dependence of susceptibility (χ_{fd}), computed from both measured parameters, indicates the approximate percentage of pedogenic ferrimagnetic grains that occur in a narrow window of ultrafine grain-sizes below the stable single domain/superparamagnetic (SD/SP) boundary (<40 nm), or the very fine clay-size fraction (Evans and Heller, 2003).

Magnetic enhancement of the loess sequences within the central Great Plains is due to magnetic mineral content of the loess at the time of accumulation and to subsequent pedogenic enhancement. Because the inherent magnetic signal of individual loess units within the region is relatively small and consistent (Johnson and Willey, 2000), enhancement is usually related to the degree of pedogenesis, though soil drainage and other factors are involved in determining the level of enhancement (Grimley and Veprakas, 2000; Evans and Heller, 2003). Recognized non-anthropogenic processes for magnetic enhancement during pedogenesis include weathering of iron-bearing minerals during cycles of wetting and drying, biomineralization of magnetite by magnetotactic bacteria, natural fires, and inorganic,

in situ formation of ultrafine-grained magnetite (Dearing *et al.*, 1996). Recently, Chen *et al.* (2005) documented inorganic formation of sub-micrometer- and nano-crystalline maghemite from larger magnetite grains in Chinese loess-derived paleosols and considered the maghemite and biogenic magnetite to be the source of susceptibility enhancement in the paleosols. Despite the tendency of χ_{fd} to exhibit high variance, elevated values in the solum of well-drained soils reflect in situ formation of ultrafine magnetite (Maher and Thompson, 1995).

Results

Stratigraphy at Buzzard's Roost consists of three major loess units and an intercalated sequence of nine buried soils. Five paleosols occur within the Loveland Loess, including the Beaver Creek Soil named herein after the Beaver Creek Formation of Reed and Dreeszen (1965), three soils within the Buzzard's Roost Paleosol Complex (BR S1, S2, and S3; Schultz and Martin, 1970), and the Sangamon soil. The overlying GCF consists of two distinct basal loess units (GCF L3 and L2), a welded soil couplet (GCF S3 and S2), an upper loess unit (GCF L1), and the weakly developed, uppermost GCF soil (GCF S1), which is overlain by the Peoria Loess.

Radiocarbon chronology

All ten radiocarbon ages obtained from the profile were in stratigraphic sequence, ranging from 38,080 cal yr BP (lower part of S3) to 28,190 cal yr BP (upper S2) and reflecting a time span of about 10 kyr (Table 2). Age control for the

Table 2. Radiocarbon ages from the road-cut profile and core at Buzzard's Roost

Depth (cm)	Stratigraphic unit	Lab. No.	¹⁴ C Age BP a	δ ¹³ C (‰)	Cal Age BP b
Profile					
300-305	upper S2	ISGS-4388	23,640±250	-22.2	28,190±280
325-330	lower S2	ISGS-4389	23,670±270	-23.4	28,220±300
356-361	upper S2	ISGS-4157	23,590±220	-23.3	28,140±250
385-390	lower S2	ISGS-4162	23,890±200	-22.2	28,450±230
420-425	lower S3	ISGS-4156	25,240±160	-20.7	30,470±240
443-448	upper S3	ISGS-4187	27,970±260	-18.1	32,700±230
465-470	lower S3	ISGS-4164	31,070±410	-18.7	36,030±470
480-485	upper S3	ISGS-4183	31,240±320	-18.3	36,220±390
505-510	mid S3	ISGS-4165	33,160±300	-16.3	38,080±560
540-545	lower S3	ISGS-4163	33,150±540	-19.2	38,080±760
Core					
186	S1	OS-51552	21,300±130	-21.1	25,580±220
587.5	S3	OS-51553	33,500±320	-21.0	38,370±640
681	S3	OS-51554	31,600±230	-22.3	36,620±220

a assays on total soil organic carbon (SOC)

b calibrated with the Fairbanks0805 calibration curve (Fairbanks *et al.*, 2005)

GCF was supplemented with three AMS radiocarbon ages from SOC within the core (Table 2). The uppermost of these, dating to 25,580 cal yr BP, reflects the age of a weak terminal A horizon (S1) capping the GCF, which was undated in the profile. The other two ages were intended to date lower boundaries of the lower two GCF loess units, below the soil ages obtained in the profile. While the first of these was stratigraphically consistent, the lowermost one, at 36,620 cal yr BP, was too young; the anomalous age was not unexpected because of extreme bioturbation evident in the lower loess unit, a common problem in the GCF and often undetectable, especially in a core.

Road-cut profile

The profile prepared in the road-cut exposure provided access to most of the GCF including GCF L2, S3, S2, L1, and S1. Stable carbon isotope values at the base of the profile indicate an abrupt upward increase in the C_4 component (from -21.3‰) within uppermost GCF L2 (Figure 4). Isotope values rise within S3 to a maximum of -17‰, signifying a C_4 -dominated grassland about 38,000 cal yr BP. As S3 development continued, C_3 plants gradually regained importance until, near the S3/S2 transition, $\delta^{13}C$ stabilized around -21‰ for 1-2 kyr (Figure 5). A further increase in C_3 plant composition is recorded within S2, where $\delta^{13}C$ drops to about -23‰. Only minor resurgences of C_4 plant input are recorded in the upper part of S2 and in GCF L1. The two samples from S1 suggest a C_3 -dominated environment

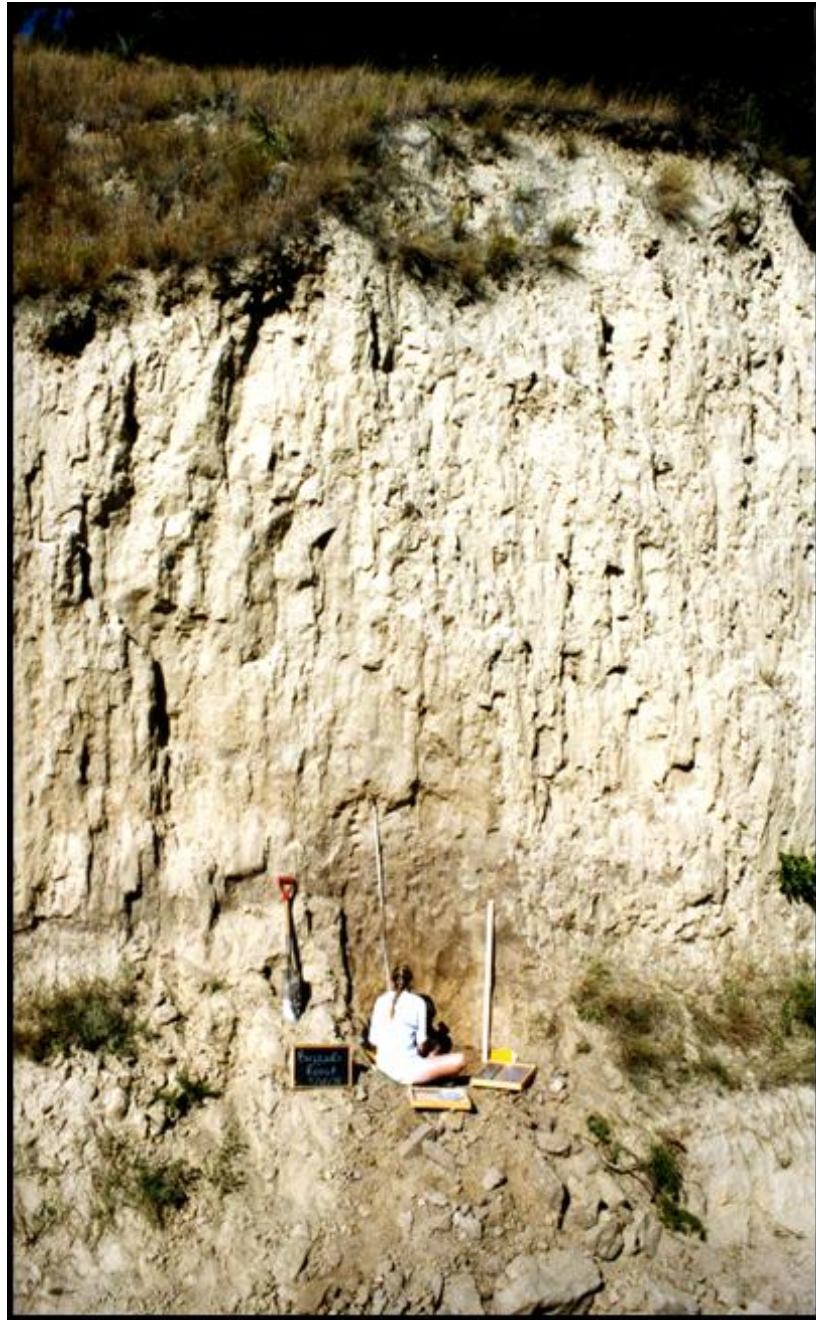


Figure 4. Sampling of the Gilman Canyon Formation profile in the road cut at Buzzard's Roost. The core site was 10m upslope (to the left).

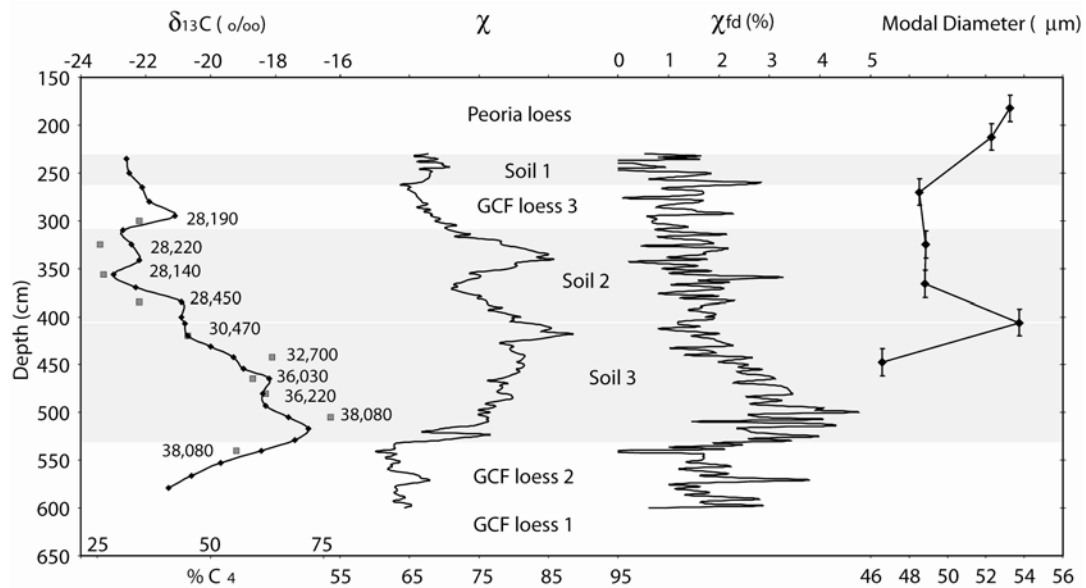


Figure 5. Stratigraphy, radiocarbon ages, $\delta^{13}\text{C}$ data, magnetic susceptibility (X , X_{fd}), and particle size distribution for the road-cut profile. Open box symbols adjacent to radiocarbon ages represent the $\delta^{13}\text{C}$ values produced by the radiocarbon laboratory to correct the ages for isotopic fractionation (Table 1). Horizontal lines define upper boundaries of loess units and soils. Vertical bars on particle size distribution data points indicate sample depth intervals.

(-22.5‰). $\delta^{13}\text{C}$ values reported with radiocarbon ages generally agree with the more detailed measurements shown in Figure 5.

Measurement of magnetic susceptibility (χ) from the profile revealed three spikes characterizing the Gilman Canyon soil complex between about 530 and 230 cm (Figure 5). The broad, lowermost spike, extending from about 530 to 375 cm, is associated with gradually increasing susceptibility through all of S3 and sharply decreasing susceptibility through the BC horizon of S2. The lower spike's peak of $89 \cdot 10^{-8} \text{ m}^3/\text{kg}$ occurs near the S3/S2 boundary (*c* 420 cm). A narrower peak ($\leq 85 \cdot 10^{-8} \text{ m}^3/\text{kg}$) from 360-310 cm occurs in the A1 and A2 horizons of S2. The minor uppermost spike starting at 260 cm and continuing to the top of the recorded profile (230 cm) is related to weak pedogenesis that produced S1 in the uppermost GCF prior to the onset of rapid Peoria Loess influx. Within the GCF L2, frequency dependence of susceptibility (χ_{fd}) is in the 2% to 1.5% range, but within S3 values rise sharply to about 3.5% within the lower part of the soil (*c* 38 ka) (Figure 5). Values of χ_{fd} decline gradually in the upper part of the S3 solum, back to a low of about 1.4% at the S3/S2 transition (*c* 30 ka). Values slowly decline upward within S2, from about 1.8% down to less than 1.0%. L3 and S1 appear to have little or no χ_{fd} enhancement above what is common for Peoria Loess.

As is typical for loess-derived sediments and soils, particle size distribution displayed only minor variation within the interval sampled. Modal diameter is smallest (46.6 μm) within the A horizon of S3 (*c* 450 cm), and increased slightly to a maximum (53.7 μm) at about 400 cm, when the sedimentation rate may have been

elevated briefly between S3 and S2 (Figure 5). The modal particle size decreased to (48.7 μm) within S2 and into GCF L1 from about 375-275 cm, before increasing (53.3 μm) within the basal Peoria Loess (< 220 cm).

Core

Almost 19 m, including the lowermost Peoria Loess, entire GCF, and 10 m of Loveland Loess were recovered in the core (Figure 6). Stable carbon isotope ratios from S3 indicate an elevated C_4 grass component very similar to that inferred from the roadcut profile. Likewise, S2 reflects a shift to a more C_3 -dominated plant community. Despite a slight rise in isotope values within S1, C_3 dominance persisted. Below the S3 peak, there are two single data-points which likely represent *Spermophilus krotovinas* penetrated by the core and filled with S3 material from above. Increased $\delta^{13}\text{C}$ within the Sangamon Soil suggest that, like GCF S3, a significant C_4 component characterized the plant community during pedogenesis. Similarly, the Buzzard Roost soil-forming periods below the Sangamon Soil exhibit marked increases in $\delta^{13}\text{C}$ values.

Magnetic susceptibility displays a major feature at about 700 cm, which consists of a decrease in the background signal from that of the Loveland Loess to that of the GCF and Peoria Loess (Figure 6). This shift likely represents a change in loess provenance, though other compositional studies will be needed to test this hypothesis. Variation in χ of the Loveland Loess relates largely to enhancement through pedogenesis and to dilution from carbonate accumulation. Due to the

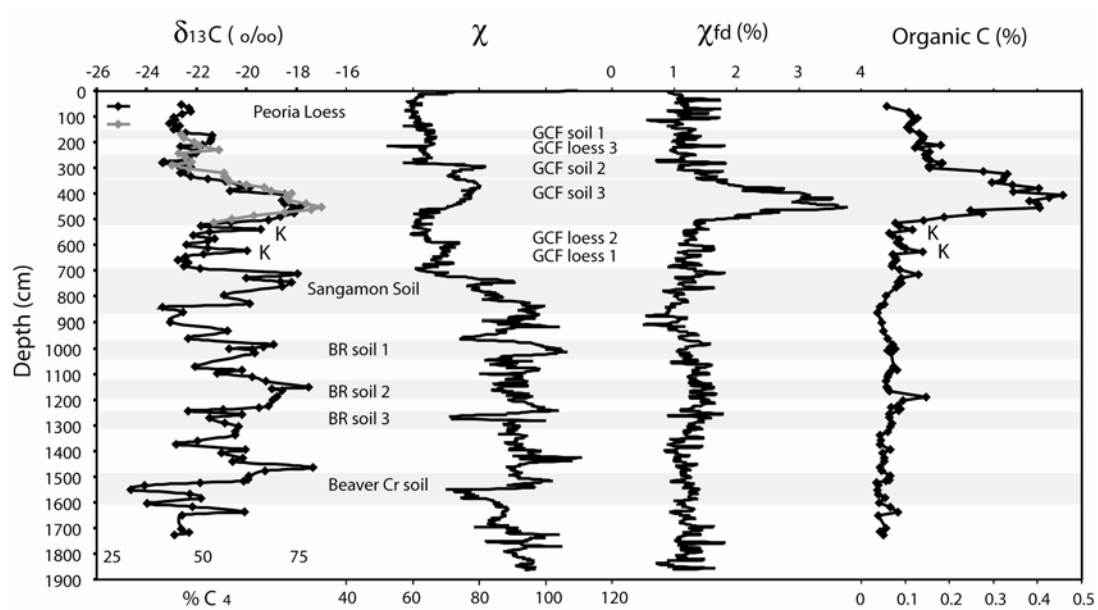


Figure 6. Stratigraphy, $\delta^{13}\text{C}$ values, magnetic susceptibility (χ , χ_{fd}), and organic carbon derived from the core. K indicates krotovinas.

interaction of these two processes, and possibly other unrecognized factors, χ does not clearly define the soils below the Sangamon Soil, with the exception of Buzzard's Roost S1 and possibly the Beaver Creek soil. Sharp drops in susceptibility occurring at about 1550, 1270, and 950 cm illustrate the carbonate dilution effect.

On the basis of the magnetic signal, color shift, and structural differences, two distinct loess units (L1 and L2) were recognized within the GCF beneath S3. Soils 3 and 2 display the multimodal χ signal exhibited by profile data. Above S2 is a third GCF loess (L3), overlain by the weakly developed S1. The upper 1.5 m of the core is Peoria Loess, with a surface spike reflecting disturbance from the roadbed. As in the profile, χ_{fd} peaks within the lower part of the GCF S3 and exhibits a low-amplitude rise coincident with the Buzzard's Roost paleosols (*c* 1370-920 cm).

Organic carbon content, determined with the elemental analyzer (Figure 6), displays a pronounced peak within S3 and S2. As in the $\delta^{13}\text{C}$ data, krotovina-derived spikes appear below S3, further confirming that origin of the fill is S2 or S3. The two minor peaks below are those of the Sangamon Soil and Buzzard's Roost S2. Decline in organic carbon content above GCF S2 reflects increased flux of sediment associated with the waning stage of the GCF and transition into Peoria Loess.

Carbon and nitrogen data from the combustion analyzer provide additional information on carbonate content and portray variability of nitrogen and its ratio to organic carbon (Figure 7). While carbon is mainly inorganic within the Loveland Loess, the GCF is leached to the bottom of its basal loess where carbonate has accumulated immediately above the Sangamon Soil. Organic carbon content

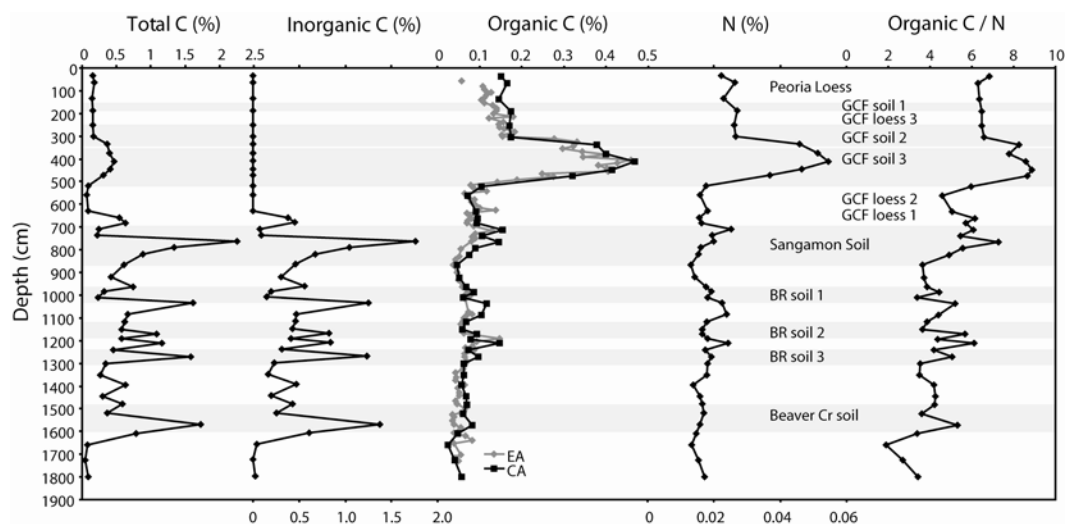


Figure 7. Carbon (total, inorganic, and organic), total nitrogen, and organic C/N data from the core. Soil designations are S (Sangamon), BR 1-3 (Buzzard's Roost Paleosols), and BC (Beaver Creek soil). Inorganic and organic carbon percentages are those of the total carbon.

highlights the GCF S3 and S2, but not the S1. Below the GCF, inorganic carbon spikes within carbonate-rich horizons and to a lesser extent immediately above the various intra-Illinoian soils. The Sangamon Soil has the most pronounced carbonate accumulations. Organic carbon content conforms well to that derived from the elemental analyzer, especially given the application of two different analytical techniques to separate suites of core samples.

Total nitrogen has a vertical distribution similar to that of organic carbon, but when expressed as a ratio with organic carbon, most soils recognized within the core become better defined (Figure 7). While C/N ratios of modern soils are typically between 10 and 12 (Stevenson and Cole, 1999), all C/N ratios from the core are less than 10 and are especially low in the older soils below the Sangamon Soil. This observation may be explained by ongoing alteration of soil organic matter after burial. In surface soils, the C/N ratio decreases rapidly as raw plant material is converted to humus (Stevenson and Cole, 1999). After burial, input of plant material ceases, but slow decomposition of carbon may continue, further concentrating nitrogen and decreasing the C/N ratio.

Discussion

Loess Flux

At Buzzard's Roost, age determinations indicate that rates of soil formation and associated loess accumulation were variable throughout development of the GCF (Figure 8). Assuming the lowermost radiocarbon age is an underestimation, an age of

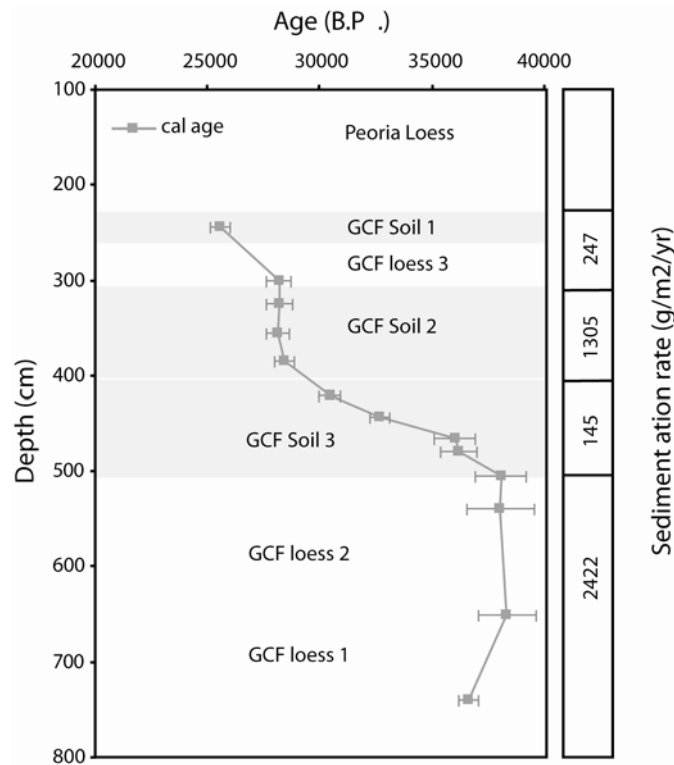


Figure 8. Composite age-depth relationship defined with numeric ages from the road-cut profile and the core. Age data are from Table 2, and error bars on radiocarbon ages represent 2σ. Depths of the AMS ages from the core have been scaled to those of the road-cut profile.

38,370 cal yr BP on GCF L2 suggests that GCF L3 began accumulating by 40 ka or earlier, indicating an accumulation rate of about 1.7 mm/yr for the two loess units. Age distribution within S3 indicates slower loess accumulation (*c* 0.1 mm/yr) after about 38.5 ka, suggesting a reduced atmospheric dust load due to more vegetation cover in the source region and/or reduced wind velocities. Accumulation rates increased substantially (*c* 0.9 mm/yr) during formation of S2 (*c* 30 ka) and then slowed to about 0.2 mm/yr between about 28.6 ka and 25.5 ka during deposition of GCF L1. Because of compaction and other diagenetic processes, mass accumulation rate (MAR) may provide a better perspective on influx of loess. Using 1.45×10^6 g/m³ as a representative bulk density of loess (Bettis *et al.*, 2003b), MARs for GCF L3 and L2, S3, S2, and L1 were about 2422, 145, 1305, and 247 g/m²/yr, respectively. Even the highest MAR for the GCF at Buzzard's Roost is well below the maximum MAR of about 11,500 g/m²/yr attained by Peoria Loess within the region (Roberts *et al.*, 2003).

Local paleovegetation and inferred climate

Presence of krotovinas and skeletal remains of ground squirrel, prairie dog, and other rodents with a grassland affinity indicate a prairie environment for the GCF soil forming intervals at Buzzard's Roost and elsewhere in central and western Nebraska (Schultz 1934, Schultz and Stout, 1948, Schultz *et al.*, 1951, Tobin, 2004a, 2004b). The association is particularly well expressed by the C₄-dominated S3 in that most krotovina fill appears to originate from within this soil. Activity of rodents

occupying S3, and to a lesser extent S2, was sufficiently intense that Tobin (2004a; 2004b) designated these soils and the Sangamon Soil-Loveland Loess as the upper and lower *Citellus* zones, respectively. Frankel (1956) interpreted mollusca in the Gilman Canyon soil at Buzzard's Roost as indicating a grassland environment with occasional and severe drought.

After 38 ka, $\delta^{13}\text{C}$ values within S3 peaked at -17.0‰ (profile) and -17.8‰ (core), reflecting a C_4 contribution of 71% and 66% to the SOC, respectively (Figures 5, 6). The $\delta^{13}\text{C}$ value of -16.3‰ (76% C_4) derived for the associated radiocarbon sample (Table 2) also indicates C_4 dominance. Application of the relationship between modern soil $\delta^{13}\text{C}$ and mean July temperature today illustrated by Fredlund and Tieszen (1997a) and quantified by Johnson and Willey (2000) suggests a mean July temperature of 25.6°C from a $\delta^{13}\text{C}$ value of -17‰. Presently, mean July temperature for the SW Division of Nebraska is 23.8°C (NOAA, n.d.), about 2°C cooler than the peak temperature inferred for Gilman Canyon time at Buzzard's Roost. As an approximate modern analog, the city of Hays in western Kansas has a mean July temperature of about 26°C and is in an area dominated by two C_4 grass genera, grama grass (*Bouteloua* spp.) and bluestem (*Andropogon* sp.). During much of GCF development, however, mean July temperatures were slightly cooler than during S3 pedogenesis and more like those of the present-day region. Lower isotopic values from S2 indicate a cool-season C_3 grass community rather than one including C_3 arboreal species because 30-40% of the SOC was derived from C_4 plants, and prairie fauna (ground squirrels and prairie dogs) were still inhabiting the area.

Average $\delta^{13}\text{C}$ values for the upper GCF (S2, L1, and S1) indicate a mean July temperature around 18.5°C (*c* 5°C lower than at present), similar to that of Regina, Saskatchewan.

Regional environmental records

Stable carbon isotope data from elsewhere in the region provide a similar but less detailed image of Middle Wisconsinan plant communities. At the Last Chance site (Figure 3), data indicate a mixed community with a bias toward C_3 plants, whereas at Beecher Island data suggest a mixed community without any obvious photosynthetic bias (Muhs *et al.*, 1999a). Although the apparent reduction in C_4 influence at these localities in Colorado, relative to Buzzard's Roost, may be due to insufficient resolution of plant community changes over time, it may also be attributable to the presence of western wheatgrass (*Agropyron smithii*) and other C_3 grasses and shrubs tolerant of drought conditions.

At the Eustis Ash Pit in southwestern Nebraska, Fredlund *et al.* (1985) reported a peak in the relative abundance of C_4 phytolith forms within the GCF soil zone, as well as an aberrant peak of panicoid phytolith forms within the GCF-to-Peoria Loess transition. Pre-Illinoian and Illinoian soils at the Eustis locality were dominated by C_4 phytolith forms, supporting the $\delta^{13}\text{C}$ -based interpretation from the Buzzard's Roost core of increased C_4 grass influence during Sangamon and earlier Illinoian soil-forming periods.

Environmental records for the mid-continent that encompass the Middle Wisconsinan, have been constructed with multiple proxies, notably fossil pollen sequences and speleothem stable isotope records. Collectively, regional data indicate climate cooling at 55-50 ka, which may have been a prelude to the beginning of Gilman Canyon loess deposition, and at 40-30 ka, which coincides with S3 pedogenesis. Fossil pollen from Cheyenne Bottoms, a wetland in central Kansas, suggest persistent shallow water levels in the basin and an upland characterized by mesic grassland-sage steppe community with limited populations of arboreal species from the start of the record at about 34 ka until about 29 ka (Fredlund, 1995). At Muscotah and Arrington marshes in northeastern Kansas, fossil pollen immediately prior to 33 ka reflects an open or savanna type of vegetation prior to a Late Wisconsinan expansion of spruce (Grüger, 1973). Fossil pollen from Boney Springs, one of the mastodon-bearing springs of western Missouri, suggests an open pine parkland before spruce invasion about 31.6 ka (King, 1973).

High resolution $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data from speleothem RC2 collected at Reed's Cave, South Dakota point to cooling starting 50-45 ka (54-50 ka in speleothem RC20) and a shift from prairie to forest cover, followed by warming and a shift toward C_4 grass about 35 ka (Serefiddin *et al.*, 2004). Stable isotope data ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) from speleothems of Crevice Cave, southeastern Missouri record a major cooling trend around 55 ka, when dominant vegetation shifted from grassland to forest (Dorale *et al.*, 1998). Oxygen isotope data for Crevice Cave indicate a subsequent increase in temperature of about 2°C around 37 ka.

The cooler, more mesic environment of the Late Wisconsinan has been documented within the region using mollusk, botanical macrofossil, and isotope data. At Buzzard's Roost, Frankel (1956) reported boreal snails extending upward from the top of the GCF. Land snail taxa, common to boreal regions were reported for Bignell Hill (Leonard, 1951), Eustis ash pit (Rousseau and Kukla, 1994), Harlan County Lake (Wells and Stewart, 1987), and other locations within Nebraska and Kansas (Leonard, 1952; Wells and Stewart, 1987). At Harlan County Lake, Wells and Stewart (1987) dated white spruce (*Picea glauca*) needles to about 17 ka, and Martin (1993) reported ages of 23-25 ka from lenses of spruce charcoal in the lower Peoria Loess. In eastern Colorado, faunal and isotopic data from the late Wisconsinan suggest a cool grassland environment like the northern Great Plains of today (Muhs *et al.*, 1999b).

Global paleoclimatic connections

Heinrich events and other oscillations have been identified in continental records from North America (Grimm *et al.*, 1993; Clark and Bartlein, 1995; Zic *et al.*, 2002), as well as in those from other continents (Wang *et al.*, 2001). If Heinrich Events coincided with weakening of North Atlantic thermohaline circulation, this could have reduced the supply of warm water to the Gulf during Heinrich events (Bond *et al.*, 1992), which in turn could have reduced the flux of moist warm air into the mid-continent (Grimm *et al.*, 1993). Thus, it is plausible that Heinrich events may be recorded as periods of limited soil formation in the loess stratigraphy of the central Great Plains.

Bond cycles culminating in Heinrich events (Bond *et al.*, 1993; 1997) coincide with times of both loess accumulation (cycle defined by H5 to H4) and soil formation (H4 to H3) (Figure 9a). The H3 to H2 Bond cycle includes times of both loess accumulation and minor soil development at Buzzard's Roost (Figure 8a). Though it is difficult to suggest causation, the strongest expression of pedogenesis and C₄ dominance, the S3, is bounded temporally by the H4 (*c* 37 ka) at its inception and the H3 (*c* 29 ka) at the transition to the more mesic S2 (Bond and Lotti, 1995).

Comparison of the loess record at Buzzard's Roost with summer insolation curves suggests a much more straightforward relationship. The height of effective soil formation within the GCF was reached in S3 at about 35 ka, a time when June insolation was at its highest level in the last 50 kyr (Figure 9b). This coincided with peak C₄ plant abundance and with warm conditions inferred from speleothems elsewhere in central North America (Dorale *et al.*, 1998; Serefiddin *et al.*, 2004).

It is noteworthy that development of the prominent Brady Soil (Schultz and Stout, 1948; Johnson and Willey, 2000) in the central Great Plains also coincides in part with peak summer insolation around 12-10 ka. Stable carbon isotopes from Brady Soil SOC indicate a much greater C₄ abundance than that within the underlying loess (Johnson and Willey, 2000; Feggestad *et al.*, 2005), similar to the GCF S3 at Buzzard's Roost. Overall, there is strong evidence for insolation forcing of climate in the loess record of the central Great Plains.

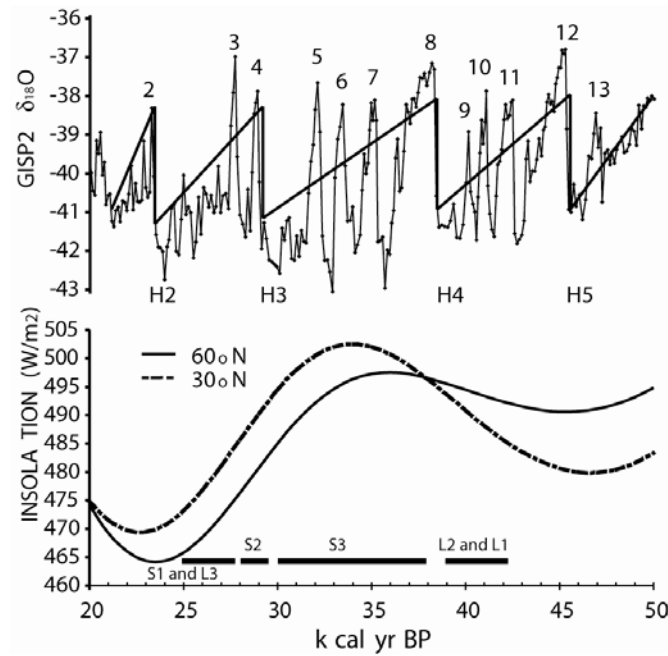


Figure 9. (a) Correlations among millennial-scale oscillations in GISP2 $\delta^{18}\text{O}$, June insolation and timing of Gilman Canyon loess deposition and soil formation. Dansgaard-Oeschger oscillation events (interstadial numbers) are from Bond and others (1993), Heinrich event chronology from Bond and others (1993, 1997), GISP2 $\delta^{18}\text{O}$ values from Stuiver and Grootes (2000). Timing of D-O oscillation events is only approximate, given recent age revisions (*e.g.*, Genty *et al.*, 2003; Shackleton *et al.*, 2004). (b) June insolation curves for 30°N and 60°N. Buzzard's Roost is located at about 40.8°N. Insolation data are from Berger and Loutre (1991).

Conclusions

Voelker *et al.* (2002) recognized a paucity of continuous climate records for MIS 3 in the mid-continent, particularly those of a centennial scale. While the record from Buzzard's Roost does not provide this resolution, it does provide an important history of climate change that occurred within the Great Plains during that time interval. Moreover, investigation of the GCF generated new data that provide insight to the loess chronology and to the environments prevailing at the site during the Middle Wisconsin. Significant findings include the following:

- 1) deposition of the GCF spanned about 15 kyr, with a basal age of at least 40 ka and an upper age of about 25 ka (latter half of MIS 3);
- 2) the formation is informally divisible into three loess units (GCF L1, L2, and L3), the upper two of which contain three soils (GCF S1, S2, and S3);
- 3) the first period of pedogenesis (S3) was most extensive, persisting for about 8 kyr (*c* 38-30 ka) during slow loess accumulation;
- 4) based on the $\delta^{13}\text{C}$ record from SOC, the environment of the GCF began and ended with a C_3 -dominated plant community, while the main soil-forming interval was characterized by a C_4 -dominated grassland;
- 5) environmental changes recorded within the GCF appear to agree with other regional proxy records of climate for that interval;
- 6) good temporal correlation exists between deposition of the GCF and formation of its soils with the Roxana Silt and Farmdale Geosol of the Upper Mississippi River valley, respectively;

- 7) peak pedogenesis (S3) occurs temporally within a single Bond cycle (H4-H3);
- 8) most effective pedogenesis and highest C₄ plant abundance during MIS 3 coincided with a peak of summer insolation, suggesting strong insolation forcing of climate and vegetation change in the central Great Plains.

The GCF, including its multiple paleosols, at Buzzard's Roost records a remarkable climatic interlude within the Wisconsin glacial period that persisted for several thousand years. For much of the time during which GCF S3 developed, the environment of the central Great Plains appears to have been somewhat similar to that of today.

CHAPTER 3: Stable Isotope and Rock Magnetic Expression of Environmental Change at the Pleistocene-Holocene Transition in the Central Great Plains

(published as Johnson and Willey, 2000)

Abstract

The Brady Soil, a ubiquitous feature capping the Late Pleistocene Peoria Loess, records the character of vegetation change associated with dynamic climate shifts during the Pleistocene-Holocene transition (*c* 14-9 ka). Stable carbon isotope ratio analysis and two magnetic parameters were used to examine the nature of the change during this period across north-south and east-west variability. Stable isotope ratios of carbon ($\delta^{13}\text{C}$) derived from soil organic matter indicate that in the Late Pleistocene minor populations of drought-tolerant (C_4) plants were interspersed among the dominant cool-season grasses (C_3) in western Kansas and Nebraska, while C_3 grasses were accompanied by trees in the east of those states. By the early Holocene, C_4 grasses were the dominant vegetation in the west and contributed substantially to the $\delta^{13}\text{C}$ signal in the east. The magnitude of the $\delta^{13}\text{C}$ shift suggests an increase in July mean daily temperature of 8-10°C, according to existing models. Magnetic susceptibility and frequency dependence of susceptibility signal the cessation of loess deposition and development of the Brady Soil and indicate increasing intensity of weathering from west to east.

Introduction

Most of what is known of the character, magnitude, and timing of environmental change during the Pleistocene-Holocene transition in the central Great Plains comes from proxy sources such as floral and faunal remains and geomorphic records. For example, Fredlund and Jaumann (1987) compiled the palynological and paleobotanical records from the Great Plains, whereas Davis (1987) examined the mammalian record at sites in Kansas and Oklahoma. Within recent years, appreciable geomorphic research in the region has recovered records of Late Quaternary environmental change in sediments including loess (Johnson, 1993; May and Holen, 1993; Feng *et al.*, 1994; Rousseau and Kukla, 1994; Miao *et al.*, 2007; Mason *et al.*, 2008), aeolian sand (Ahlbrandt *et al.*, 1983; Muhs, 1985; Forman *et al.*, 1995; Madole, 1995; Arbogast and Johnson, 1998; Forman *et al.*, 2008; Mason *et al.*, 2004; Hanson *et al.*, 2009), alluvium (Johnson and Logan, 1990; May, 1992; Mandel, 1994, 2008), and playa-lunette systems (Arbogast, 1996; Holliday, 1997; Fredlund *et al.*, 1998; Holliday *et al.*, 2008).

Loess deposits represent a nearly continuous terrestrial sedimentary sequence, despite periodic punctuation by soil formation; Kukla (1970, 1977) portrayed their importance as equivalent to the marine sedimentary sequences. Correspondingly, loess deposits of the central Great Plains and their intercalated soils contain the potential for extracting high-resolution terrestrial records (Fredlund *et al.*, 1985; Muhs *et al.*, 1999; Muhs *et al.*, 2008). Further, these deposits offer the potential for the application of a rich array of climatic proxies, such as carbon isotopes ($\delta^{13}\text{C}$), opal

phytoliths, rock magnetism, sedimentology, and vertebrate and invertebrate paleontology. Stable carbon isotope ratios and two, room-temperature, rock magnetic parameters have been used to document environmental change at the Pleistocene-Holocene transition.

Late-Quaternary environments of the central Great Plains

Late Pleistocene

Following a warm interval between (40-25 ka) during which the Gilman Canyon Formation loess and soils developed, temperatures in the region appear to have been 5-10°C cooler, and precipitation slightly less than today (Kutzbach and Wright, 1985; Kutzbach, 1987). The Late Wisconsinan was characterized in General Circulation Models (GCMs) by a bifurcation in the polar jet in response to the North American ice sheet (Kutzbach and Wright, 1985; COHMAP, 1988; Kutzbach *et al.*, 1998). Wind flow patterns from the southerly branch resulted in northwesterly winds across the central Great Plains, which is evident in the directional expression of Pleistocene aeolian landforms (Wells, 1983; Muhs and Bettis, 2000). There is, however, a discrepancy in wind direction between the terrestrial physical evidence and the circulation predicted by the GCMs (Bartlein *et al.*, 1998; Mahowald *et al.*, 1999; Muhs and Bettis, 2000). Regionally, relict parabolic dune orientations, indicating northwesterly paleowinds, occur in the Sand Hills of Nebraska (Swinehart, 1990), northeastern Colorado (Muhs, 1985), and central Kansas (Arbogast and

Johnson, 1998). Lunettes associated with playas also reflect northwesterly flow (Arbogast, 1996; Holliday, 1997).

Loess from the Platte River of Nebraska, and to a lesser extent from other east-flowing rivers to the south, mantles much of Nebraska (Reed and Dreeszen, 1965; Reed, 1968; Ruhe, 1983) and Kansas (Frye and Leonard, 1951; Johnson, 1993; Feng *et al.*, 1994). The proxy record contained within the Late Wisconsinan loess correlates well with the Greenland ice core proxies. For example, grain-size data from the Eustis ash pit in southwestern Nebraska correspond to the dust and $\delta^{18}\text{O}$ records from the GRIP ice core, documenting the global-scale signal contained within the loess deposits (Rousseau *et al.*, 2007).

A variety of faunal remains provides proxy information about the nature of the Late Pleistocene in the central Great Plains. Leonard (1952) used loess-derived snails to subdivide Late Wisconsinan loess into zones and to deduce a boreal environment. From mollusca and associated small mammalian fauna collected at scattered sites in the region, Wells and Stewart (1987) suggested a cordilleran-boreal environment including temperate, moist broadleaf-deciduous forest vegetation containing *Populus tremuloides* (quaking aspen). Rousseau and Kukla (1994) defined three molluscan zones in Late Wisconsinan loess at the Eustis ash pit in southwestern Nebraska, and proposed a grassy steppe between about 24 ka and 16 ka, which was followed by a dry and cold grass steppe with scattered shrubs and few trees from about 16 ka to 12 ka. Collectively, regional faunal data suggest a cool, moist climate, lacking the seasonal extremes of the present (Graham, 1987; Martin and Martin, 1987).

Throughout the region, proxy data from floral material provide a similar record of cooler growing-season temperature and greater effective moisture during the Late Holocene. The pollen record from Muscotah Marsh in northeastern Kansas indicates a spruce cover persisting until about 12 ka (Grüger, 1973). From the Rosebud site on the northern edge of the Sand Hills of Nebraska, spruce pollen was the most common arboreal taxon between 13 ka and 12 ka (Watts and Wright, 1966). At sites in south-central Nebraska and north-central Kansas, Wells and Stewart (1987) uncovered *Picea* (spruce) charcoal and *Pinus flexilis* (limber pine) cones, twigs, and needle leaves (dating to about 14.5 ka), from which they interpreted taiga-like vegetation. A mixed coniferous-deciduous forest environment was reconstructed from a site in the Arkansas River valley of south-central Kansas: a peat deposit, dating to about 19 ka, contained *Picea* plant fragments and a pollen assemblage dominated by *Picea*, but containing other arboreal taxa including *Populus*, *Quercus* (oak), *Betula* (birch), and *Ulmus* (elm) (Jaumann, 1991). A pollen record from the Sanders's well site extended from about 23.7 ka to 12.8 ka and recorded aspen parkland; the exposed and more fire-prone upland environment likely excluded spruce (Fredlund and Jaumann, 1987). Conifer remains, primarily *Picea*, have been recovered from scattered sites elsewhere within the region (May, 1989; Swinehart, 1990; Johnson, 1991; Martin, 1993; May and Holen, 1993; Johnson, 1996).

Pleistocene-Holocene transition

GCM simulations indicate that by 12 ka the northern branch of the jet stream had moved south and merged with the southern branch, presumably due to a reduction in size of the ice sheet (Kutzbach, 1987; Bartlein *et al.*, 1998). Although the modeled jet stream path changed little between 12 ka and 9 ka, the intensity decreased to levels experienced today. Overall, the jet stream patterns between 12 ka and 9 ka represent the transition from the Last Glacial Maximum to the Holocene (Kutzbach, 1987; COHMAP, 1988). Beginning about 12 ka, insolation increased during northern hemisphere summers, and the influence of the ice sheet diminished, signaling an increase in seasonality (Kutzbach and Webb, 1993; Webb *et al.*, 1993). Consequently, climate and associated vegetation experienced tremendous changes during the 12-9 ka period (Jacobson *et al.*, 1987; Webb *et al.*, 1993).

At about 12 ka in the pollen record from Muscotah Marsh, *Picea* frequencies began to decline dramatically, as the spruce forests abandoned the region, with deciduous trees, grasses and forbs filling the void (Grüger, 1973). Spruce continued to decline until its disappearance from the region at about 10.5 ka; deciduous species continued to increase until about 9 ka. At Cheyenne Bottoms, a large depression in central Kansas, Fredlund (1995) documented a significant increase in *Chenopodeaceae-Amaranthaceae* (Cheno-Am) pollen after 11 ka, indicating an increase in water table variability. Grasses had become dominant in the region by 10.5 ka. From the opal phytolith record at the DB site in northeastern Kansas, Bozarth (1998) noted that the Pleistocene-Holocene transition started with an increase in the

mesic C₄ (dry, warm-season) grasses such as bluestems and grama grasses and a decrease in the C₃ (mesic, cool-season) grasses such as wheatgrass and wild rye. This trend continued through the transition but was punctuated by a brief return to cooler, more moist conditions interpreted as a response to the Younger Dryas climatic episode.

Perhaps the best paleontological locality documenting this climatic transition within the region is the North Cove site, fauna-rich spring deposits exposed in a wave-cut face on Harlan County Lake in south-central Nebraska. White spruce (*Picea glauca*) wood from the lower part of the spring dated 14.7-12.7 ka, and an organic-rich zone (12.6 ka) capping the spring contained Bison (*Bison occidentalis*) bone dating 11.4-10.1 ka (Johnson, 1989). Spring deposits were of two generations (*c* 15-10.5 ka and *c* 10 ka) and contained unique faunal assemblages. The older generation contained terrestrial and aquatic fauna with a boreal affinity suggesting spruce parkland, whereas material from the younger conveyed a somewhat warmer climate with more deciduous trees (Fredlund, 1989; Stewart, 1989).

Holocene

Beginning at 9 ka, GCM simulations portray an increased role of higher summer insolation, which was reflected in higher surface temperatures over the continent, except adjacent to the remaining ice sheet. Southern parts of the central Great Plains experienced strengthened southerly winds corresponding to an increase in cyclonic precipitation near the ice margin (Kutzbach, 1987). Summer temperatures

for 6 ka were simulated to be 2-4°C higher than today (COHMAP, 1988; Kutzbach *et al.*, 1993), with lower precipitation and stronger surficial westerly winds in the central continent. After 6 ka, the model indicates weakening westerly winds and decreased summer temperatures (Webb *et al.*, 1993).

By 9 ka, the pollen record at Muscotah Marsh was dominated by grass (Grüger, 1973). Despite unconformities and little absolute time control, a subsequent middle Holocene expansion and contraction of prairie vegetation is apparent. After 5 ka, climatic variability is suggested by the nature of the pollen record. At Cheyenne Bottoms, pollen assemblages from the Holocene are dominated by Cheno-Am types, suggesting mudflat communities associated with fluctuating water levels (Fredlund, 1995). Cheno-Am types are, however, depressed during the middle Holocene (8.5-3.7 ka), suggesting a more stable, though lower water level; an associated increase in *Ambrosia* (ragweed) may indicate that succession had reclaimed the periodically exposed mudflat areas. After 3.7 ka, Cheno-Am types rise again, suggesting renewed variability in water levels. Late-Holocene climatic fluctuations are also indicated by periodic mobilization of sand sheets and dunes within the region (Ahlbrandt *et al.*, 1983; Muhs, 1985; Swinehart, 1990; Madole, 1995; Arbogast and Johnson, 1998; Forman *et al.*, 2008; Mason *et al.*, 2004; Hanson *et al.*, 2009).

Late Quaternary stratigraphy

Five major loess units are common within the central Great Plains, including the undifferentiated pre-Illinoian pre-Loveland Loess, the Illinoian Loveland Loess,

the Middle Wisconsinan Gilman Canyon Formation loess, Late Wisconsinan Peoria Loess, and the Holocene Bignell loess (Figure 10). Regionally expressed buried soils within this interval include four or more unnamed soils in the Loveland and pre-Loveland loesses, the Sangamon Soil capping the Loveland loess, the Gilman Canyon soil, the Brady Soil of the Pleistocene/Holocene transition, and various weakly developed soils within the Bignell loess.

Pre-Loveland loess deposits, limited in outcrop expression, have been recognized, but their chronostratigraphic and lithostratigraphic relationships are poorly understood (Markewich *et al.* 1998; Bettis *et al.* 2003). Some regional information has, however, been reported. Fredlund *et al.* (1985) presented data and interpretations of opal phytolith signatures from presumed pre-Loveland strata at the Eustis ash pit in southwestern Nebraska. Feng *et al.* (1994) reported magnetic susceptibility data for pre-Loveland loess and sand on uplands adjacent to the Arkansas River in central Kansas. May *et al.* (1995) described stratigraphy (*e.g.*, pedostratigraphy) at the Elba Cut along the Fullerton Canal in west-central Nebraska, which presumably extends into the early Miocene. Most recently, Mason *et al.* (2007) identified and named the Kennard Formation, a pre-Loveland fine-grained (loess?) unit in eastern Nebraska.

The Loveland Loess, the most-widespread of the pre-Wisconsinan loess in the Midcontinent, has been recognized in Nebraska (Reed and Dreeszen, 1965) and Kansas (Frye and Leonard, 1951; Johnson, 1993). Thermoluminescence (TL) ages obtained on the Loveland Loess in the region include 163 ka, collected 1.5 m below

the Sangamon Soil at the Eustis ash pit (Maat and Johnson, 1996), and 130 ka and 136 ka from the upper Loveland Loess in northeastern Kansas (Karlstrom *et al.*, 2008). These TL ages correspond well with those derived from the Loveland Loess at the type locality in western Iowa (Forman *et al.*, 1992).

The Sangamon Soil (a.k.a. Loveland Soil, Sangamon Geosol), a strongly developed pedocomplex (Fredlund *et al.*, 1985; Morrison, 1987), is widely expressed in the Midcontinent (Muhs and Bettis, 2003). Typically this soil is relatively thick (1-2 m) with strong structural development with clay films (angular blocky to prismatic), 7.5YR hues and evidence of leaching. Overlying the Sangamon Soil is the Gilman Canyon Formation, a relatively thin (<2 m) loess unit dominated by soil development (Reed and Dreeszen, 1965). The soil within the Gilman Canyon is often sufficiently developed as to be welded to the Sangamon Soil. Ages on the soil range from greater than 40 ka to about 25 ka (Johnson *et al.*, 2007a; Muhs *et al.* 2008), while the basal age of the loess has yet to be determined.

Regionally, the Late Wisconsinan Peoria Loess is the most extensive loess in the region and often exceeds thicknesses of 10 m. It is a variably fossiliferous, but otherwise relatively homogeneous, massive silt that is light yellowish tan to buff (Frye and Leonard, 1952). Thicknesses of Peoria Loess are greatest along the southern boundary of the Sand Hills and become increasingly thinner to the south and southeast (Thorp and Smith, 1952; Ruhe, 1983).

The Brady Soil was first named by Schultz and Stout (1948) at the Bignell Hill type locality, a road cut exposure in the bluffs along the south side of the Platte

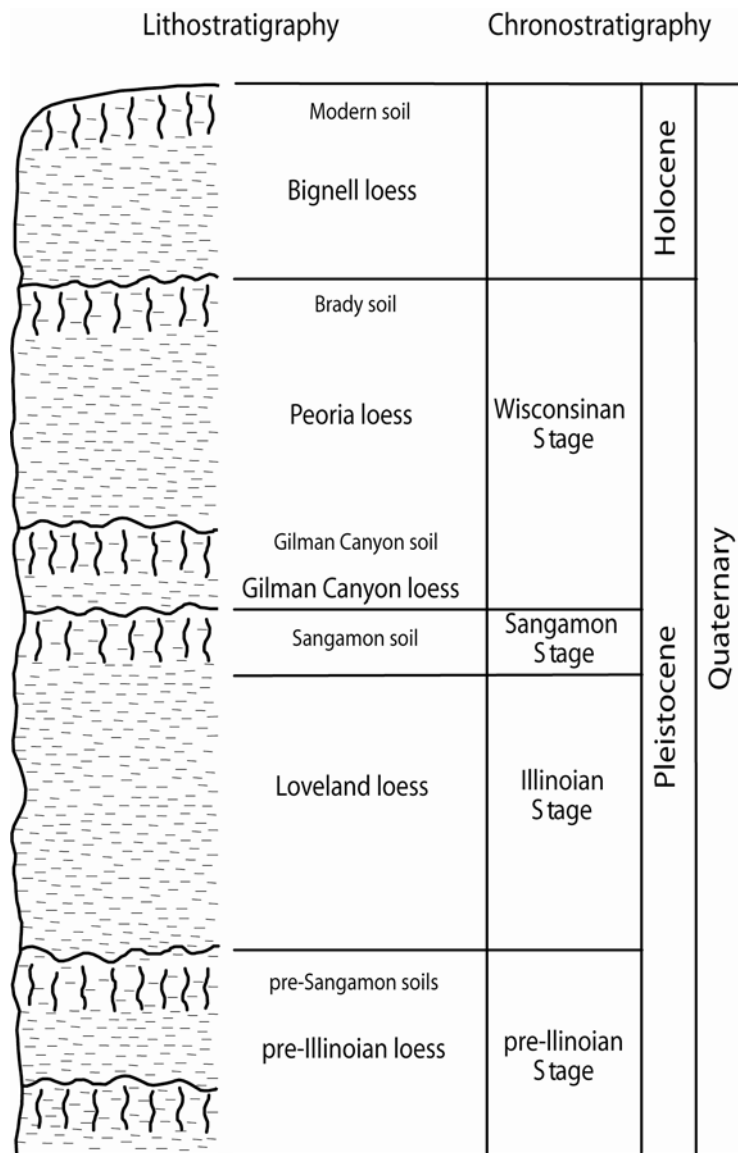


Figure 10. Composite Late Quaternary loess stratigraphy of the central Great Plains. The Sangamon Soil is about 120 kyr old; the Gilman Canyon Soil is about 40-25 kyr old; and the Brady Soil is about 13-9 kyr old. Vertical scaling varies.

River. As accumulation of the Peoria Loess diminished, Brady pedogenesis occurred to produce a generally dark gray to gray brown soil with strong textural B horizon development. Since Dreeszen (1970) first radiocarbon dated the Brady at the type locality, the soil has been dated there and at many other localities in the region (Johnson, 1993; Mason *et al.*, 2008; Muhs *et al.* 2008). The Brady Soil is not detectable unless buried beneath the Holocene Bignell loess, designated by Schultz and Stout (1945). Where not buried by Bignell loess, the Brady Soil has either become overprinted with modern soil development (Thorp *et al.*, 1951; Dreeszen, 1970) or been removed by erosion, along with the overlying Bignell loess. Multiple, weakly-developed soils within the Bignell loess indicate that the accumulation rates varied through time (Thorp *et al.*, 1951; Miao *et al.*, 2007), presumably in response to climatic variation (Mason, 1998; Mason and Kuzila, 2000; Miao *et al.*, 2007).

Study sites

The network of study sites was established roughly along two transects, one northwest to southeast and another west to east, in order to investigate paleoclimatic gradients and other aspects of variability in the upper Peoria Loess, Brady Soil, and Bignell loess (Figure 11). All sites are located on loess-mantled bluffs of river valleys or “tables,” *i.e.*, remnants of the Late Pleistocene upland loess surface. Dankworth Canyon, one of a series of large, steep canyons cut into the loess-mantled bluff on the south side of the North Platte River valley, consists of a gully face at the head of the

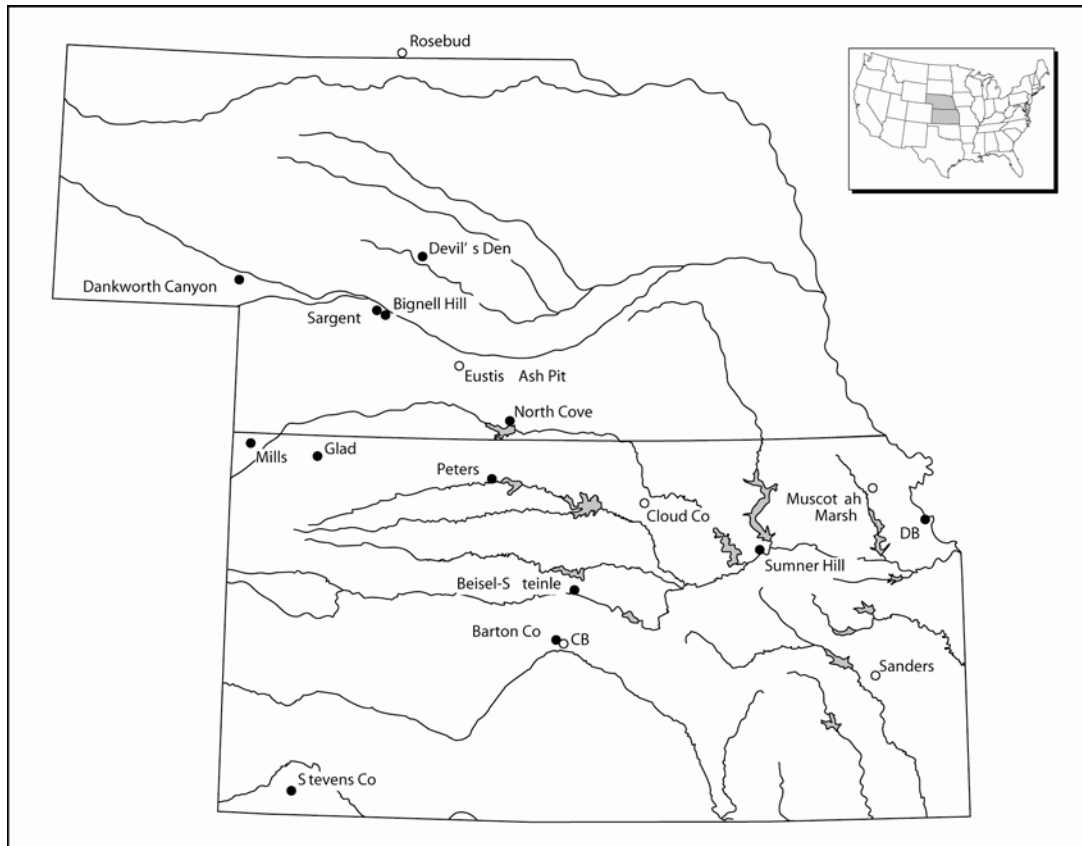


Figure 11. Sites referenced in the text and tables. Those with radiocarbon ages from the Brady Soil are represented by a solid dot.

canyon exposing the GCF up through a 3 m-thick cap of Bignell loess. The Sargent site, another gully face, was exposed by headward erosion of the Bignell Table, located on the south side of the Platte River valley; about 30 m of Peoria Loess is exposed below 3 m of Bignell loess (Figure 12). Bignell Hill, the type locality of the Brady Soil and Bignell loess, is a road cut exposure on the west end of the Bignell Table, exposing the GCF, about 51 m of Peoria Loess, and 4 m of Bignell loess. The site at Devil's Den is a face in the wall of a canyon that cuts a prominent table on the north side of the South Loup River valley and exposes about 30 m of Peoria Loess and 3.5 m of Bignell loess. Wave-cut faces in the valley walls of North Cove, a drowned tributary valley on Harlan County Lake (Republican River), expose the Sangamon soil, GCF, 8 m of Peoria Loess and about 1.5 m of Bignell loess.

The Mills site, located in a canyon draining to the North Fork of the Republican River, exposes the Sangamon soil on Pierre Shale, the GCF, approximately 23 m of Peoria Loess, and 2 m of Bignell loess. The Glad and Peters sites, road cuts along the south valley wall of Beaver Creek and north valley wall of North Fork of the Solomon River, respectively, expose the upper few meters of the Peoria Loess, the Brady Soil, and about a meter of Bignell loess. A cell of the sanitary land fill in Stevens County, Kansas, exposed the upper Peoria Loess, Brady Soil, and 1 m of Bignell loess on the upland south of the Cimarron River. Barton County sanitary land-fill cells exposed an 18 m-thick sequence of pre-Illinoian to Bignell loess (*c* 1 m thick) on the divide separating Cheyenne Bottoms to the north and the Arkansas River to the south. Stratigraphy at the Beisel-Steinle site, a road cut through



Figure 12. Exposure of the upper Peoria Loess, Brady Soil (at 6 m depth), and Bignell loess at the Sargent site. This is a west-facing view of a gully breaching the upland loess table. Refer to the 2.5m-long fence post exposed in the side of the gully for scale (arrow).

the loess-mantled bluff along north wall of the Smoky Hill River valley, includes the Sangamon soil through Bignell loess (c 1.2 m). The Sumner Hill and DB sites were 3.5 m-deep backhoe excavations on bluffs above the Kansas River and Missouri River, respectively. The former exposed about 2.3 m of Bignell loess, and the latter about 1 m.

Methods

Stable carbon isotope ratio analysis

In recent years, stable isotope ratio analysis of carbon has proven to be of use in determining past vegetation and associated climatic conditions. Analyses have been performed on pedogenic carbonate (Cerling, 1984, 1992; Cerling and Hay, 1986; Cerling *et al.*, 1989; Humphrey and Ferring, 1994), lacustrine carbonate (Humphrey and Ferring, 1994), clastic sediments (Jasper and Gagosian, 1989; Aucour *et al.*, 1994; Nordt *et al.*, 1994; Wang and Follmer, 1998), soil organic matter (Krishnamurthy *et al.*, 1982; DeLaune, 1986; Schwartz *et al.*, 1986; Guillet *et al.*, 1988; Fredlund and Tieszen, 1997a, b; Johnson *et al.*, 2008), and opal phytoliths (Kelly *et al.*, 1991, 1993; Fredlund, 1993). Recent research comparing the C₄ record from buried soils in the Great Plains with fossil pollen, marine core, and ice core data indicated that $\delta^{13}\text{C}$ data are indeed viable proxies of climate (Nordt *et al.*, 2008).

Carbon isotope fractionation occurs during photosynthesis (Smith and Epstein, 1971) as plants fix atmospheric carbon along one of three pathways: C₃ (Calvin-Benson), C₄ (Hatch-slack), and CAM (Crassulacean). The last pathway is not

relevant here, as it is a desert adaptation that selectively utilizes both photosynthetic pathways. The carbon isotopic composition ($\delta^{13}\text{C}$) of the plant material is highly correlated with the type of photosynthetic pathway used by the plant (Deines, 1980). Vascular plants segregate into two groups on the basis of their isotopic composition, or $\delta^{13}\text{C}$ value, expressed in parts per mil with respect to the PDB standard (derived from Cretaceous belemnite fossils of the PeeDee Formation). Warm-, dry-season adapted plants (C_4) have an average $\delta^{13}\text{C}$ value of -13‰ , while cool-, moist-season adapted plants (C_3) average -27‰ (Smith, 1972; O'Leary, 1988).

Soil organic matter-derived $\delta^{13}\text{C}$ values approximate the carbon fixation path of the associated vegetation, and therefore reflect the climatic environment (Tieszen and Boutton, 1989). Photosynthetic pathways of Great Plains grasses are determined primarily by the temperature regime (Teeri and Stowe, 1978; Burke *et al.*, 1991; Epstein *et al.*, 1997). While the C_4 photosynthetic pathway is associated primarily with grasses, the C_3 photosynthetic pathway is used by both trees and some grasses.

The carbon isotopic signal is imparted to the soil organic matter through decomposition of plant material. Although factors such as plant landscape position, plant life span, nitrogen nutrition, organic matter cycling, and differential decomposition and preservation can have an effect on the resulting carbon isotope composition of soil organic matter, these factors are relatively minor when using $\delta^{13}\text{C}$ values as proxy indicators of climate (Boutton, 1996). Analyses were conducted on total humates from bulk samples in order to assess the broad scale patterns of change and to minimize the impact of any given factor.

Rock magnetic analyses

The application of analytical magnetic techniques has traditionally been limited to rocks and used to determine change in the intensity and direction of the earth's magnetic field (paleomagnetism) and the origins and qualities of rock and mineral magnetism (rock magnetism). In the last two decades, however, rock magnetic analyses have been applied to environmental problems, creating the field of environmental magnetism (Thompson *et al.*, 1980; Thompson and Oldfield, 1986; Verosub and Roberts, 1995; Evans and Heller, 2003). The most common parameter employed is low-field, reversible susceptibility (χ), a nondestructive, room-temperature measure of a remnant magnetism in response to a weak external field. As a concentration-dependent parameter, susceptibility depends on the amount of magnetic material present (typically magnetite and maghemite), but can also be affected to some extent by other factors such as magnetic grain size, shape, and mineralogy. Susceptibility is likewise a function of the frequency at which it is measured. The smallest magnetic grains (*c* 20 nm) are magnetically unstable at lower frequencies and contribute to the susceptibility, whereas at higher frequencies they become stable and add less to the susceptibility. Frequency dependence of susceptibility (χ_{fd}), expressed as a percent, increases with the concentration of ultrafine magnetic material.

Many rock magnetic studies have been reported for the loess sequences of China (Kukla, 1987; An *et al.*, 1991; Maher and Thompson, 1991; Hunt *et al.*, 1995; An and Porter, 1997; Han and Jiang, 1999), and to a lesser extent for other parts of

the world, such as North America (Begét *et al.*, 1990; Hayward and Lowell, 1993; Feng *et al.*, 1994; Grimley *et al.*, 1998), Europe (Kukla, 1975, 1977), and New Zealand (Pillans and Wright, 1990). Some studies have yielded paleoclimate proxy information that has exhibited a high correlation with the marine $\delta^{18}\text{O}$ data (Heller and Liu, 1984; Kukla *et al.*, 1988; Hovan *et al.*, 1989; Liu and Ding, 1993; Banerjee, 1995). In most areas, buried soils within the loess sequences exhibit elevated susceptibility, interpreted as warmer interglacials or interstadials. Although a number of causes have been proposed for the magnetic enhancement of buried soils (Eyre and Shaw, 1994), most investigators attribute it to the *in situ* development of magnetic minerals during pedogenesis (Maher and Taylor, 1988; Zhou *et al.*, 1990; Heller *et al.*, 1993).

Susceptibility and other magnetic parameters have been used to characterize soil development recorded within loess sequences of the Midcontinent (Feng *et al.*, 1994; Grimley *et al.*, 1998; Wang *et al.*, 2009) and elsewhere (Le Borgne, 1955; Neumeister and Peschel, 1968; Thompson and Oldfield, 1986; Vadyunina and Babanin, 1972). Soils become magnetically enhanced through a number of processes associated with pedogenesis and plant cover. The formation of secondary ferrimagnetic minerals is typically the main source of the increased susceptibility. Magnetite is usually the dominant ferrimagnetic mineral, although maghemite and, to a lesser extent, titanomagnetite and titanomaghemite may also be present (Longworth and Tite, 1977; Longworth *et al.*, 1979). Most of the ferrimagnetic minerals are formed biologically within magnetotactic bacteria, either intra-cellularly (Fassbinder

et al., 1990) or extra-cellularly (Lovley *et al.*, 1987; Stolz, 1992). Some ultrafine-grained magnetite may form non-biologically through precipitation from iron solutions in an oxidizing environment (Taylor *et al.*, 1987; Maher and Taylor, 1988, 1989). Fire may also result in the formation of fine-grained ferrimagnetic minerals (magnetite or maghemite) in soils (Van der Marel, 1951; Le Borgne, 1960; Longworth *et al.*, 1979; Rummery *et al.*, 1979; Thompson and Oldfield, 1986; McClean and Kean, 1993). Even plants appear to have the potential to contribute to soil susceptibility by the deposition of phytoferritin (magnetite) upon burning of the wood and other plant material (McClean and Kean, 1993; McClean *et al.*, 1994).

Field and laboratory procedures

All samples were collected from existing exposures that had been freshened through backhoe or hand excavation, or from 5.5 cm-diameter cores captured in plastic liners. All radiocarbon and $^{13}\text{C}/^{12}\text{C}$ samples were prepared by sieving to remove fine sand and coarser material, treating with concentrated HCl to remove carbonates, drying at 100°C, and pulverizing. Radiocarbon samples were submitted to the University of Texas and Illinois State Geological Survey radiocarbon laboratories, and stable carbon isotope samples were submitted to Geochron Laboratories.

Samples for magnetic analysis were collected in plastic, cubic sample boxes (8 cm³) through insertion into the exposure profile or the core in a contiguous fashion, resulting in about 40 samples per meter. After drying, samples were measured with a Bartington magnetic susceptibility meter and dual-frequency sensor. Susceptibility,

recorded in SI units ($10^{-8} \text{ m}^3/\text{kg}$), was normalized by mass, and frequency dependence was computed from measurements at 0.465 kHz and 4.65 kHz.

Results and discussion

Radiocarbon ages

Age data from thirteen sites in the region (Table 3) were obtained from samples collected in the upper and lower 5 cm of the Brady Soil A horizons, except where there was evidence of welding with the modern surface soil. The lower and upper A horizon ages are taken to be approximations of the inception and termination of pedogenesis, respectively. The cumultic character of these loessal soils makes this a viable supposition. Erosion of the A horizon prior to burial can potentially increase the apparent terminal age, however, and is not always detectable.

Variation among the upper and lower A horizon ages is appreciable and ranges overlap, with lower ages ranging from $13,680 \pm 100$ cal yr BP to $11,150 \pm 100$ cal yr BP, and upper ages from $12,020 \pm 300$ cal yr BP to $8,780 \pm 140$ cal yr BP. Duration of pedogenesis, as measured by the difference between paired ages, ranges from $3,190 \pm 130$ cal yr BP at Dankworth Canyon to $1,240 \pm 300$ cal yr BP at Barton County land fill, for an average of $c 2,000$ yr.

Given the time-transgressive nature of climate and associated vegetation changes (Webb *et al.*, 1993; Shuman *et al.*, 2002), lower and upper A horizon ages from the Brady Soil were compared with latitude and longitude to identify any spatial component to the onset and termination of Brady genesis (Figure 13). The limited

Table 3. Radiocarbon ages from the Brady Soil

Site	North latitude	West Longitude	Depth (cm)	Sample ID
Devil's Den	41.5	100.2	426	ISGS-4102
			465	ISGS-4108
Dankworth Canyon	41.3	102.0	337	ISGS-4106
			409	ISGS-4099
Sargent	41.0	101.5	450	Tx-6934
			528	Tx-7496
Bignell Hill, a	41.0	101.5	185	Tx-7425
			220	Tx-7358
North Cove, a,b,c	40.1	99.3	155	Tx-6320
			310	Tx-6321
Mills	39.9	101.8	230	ISGS-3929
			287	ISGS-3915
Glad	39.8	101.0	118	ISGS-4097
			163	ISGS-4098
Peters, a,d	39.7	99.4	135	Tx-6626
			185	Tx-6627
DB, e	39.4	94.9	205	ISGS-3568
Sumner Hill	39.2	96.8	200	ISGS-3142
Beisel-Steinle, f,g	38.8	98.5	245	Tx-8186
Barton Co. Landfill, a,h	38.4	98.7	100	Tx-7045
			140	Tx-7046
Stevens Co. Landfill	37.2	101.3	118	Tx-8541

a Johnson, 1993; b Johnson, 1989; c Martin, 1990; d Johnson, 1990; e Logan, 1998

f Diekmeyer, 1994; g Arbogast and Johnson, 1996; h Feng, 1991; i Fairbanks *et al.*, 2005

Table 3. Continued

Site	Depth (cm)	¹⁴ C Age yr BP	δ ¹³ C (‰)	Calendar Age (yr BP) i
Devil's Den	426	9,010±70	-18.2	10,190±70
	465	10,230±80	-19.9	11,960±150
Dankworth Canyon	337	8,520±70	-18.9	9,510±40
	409	10,780±110	-20.4	12,700±90
Sargent	450	9,780±150	-16.2	11,190±200
	528	10,640±70	-20.1	12,590±70
Bignell Hill	185	9,240±110	-17.4	10,410±150
	220	10,670±130	-19.3	12,610±130
North Cove	155	10,270±160	-15.5	12,020±300
	310	11,530±150	-18.9	13,390±160
Mills	230	7,940±70	-16.5	8,780±140
	287	9,720±80	-17.5	11,150±100
Glad	118	9,630±70	-17.4	11,020±150
	163	11,800±110	-20.0	13,650±100
Peters	135	8,850±140	-18.8	9,930±240
	185	10,050±160	-17.5	11,600±310
DB	205	10,510±190	-23.9	12,400±270
Sumner Hill	200	11,230±310	-16.8	13,090±300
Beisel-Steinle	245	11,350±110	-20.1	13,200±130
Barton Co. Landfill	100	9,820±110	-18.6	11,230±120
	140	10,550±150	-19.0	12,470±200
Stevens Co. Landfill	118	11,240±70	-19.2	13,080±70

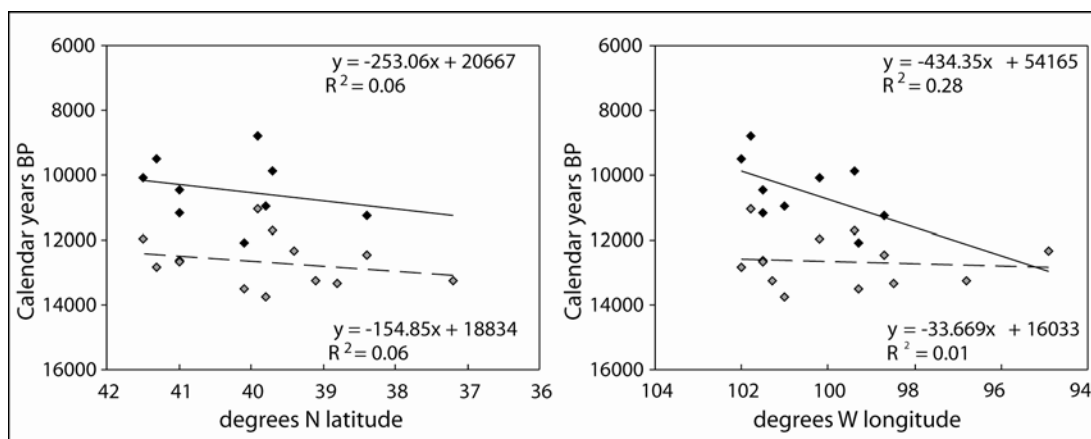


Figure 13. Correlation of upper and lower Brady Soil ages with latitude and longitude. Solid regression lines represent the upper Brady ages, and dashed the lower Brady ages.

data suggest that no discernable north-south trend exists for either the upper or lower Brady ages. Across the east-west transect there is no spatial trend for basal Brady age, but there is a weak trend suggesting that Brady Soil development may have given way to Bignell loess deposition earlier in the east. It is also possible that younger upper Brady ages went unsampled, as they tended to be welded to the modern surface soil. Additional data will add clarity as they become available.

Carbon isotope analysis

Carbon isotope curves (Figures 14-16), assembled for three sites (DB, Beisel-Steinle, Mills) along an east-west transect across Kansas, portray the change in vegetation regime from the Late Pleistocene (uppermost Peoria Loess and lowermost Brady Soil) to the early Holocene (uppermost Brady Soil). At DB, $\delta^{13}\text{C}$ values average -24.1‰ and -18.3‰ for the Peoria Loess and upper Brady Soil, respectively, for a 5.8‰ shift. The decline in values and lack of data near the surface reflect historical disturbance. A value of -22.8‰ is representative of the Peoria Loess at Beisel-Steinle, whereas -15.8‰ characterizes the upper Brady Soil, for a shift of 7.0‰. For the Mills site, Peoria Loess registers about -21.3‰ and the upper Brady about -13.7‰, for a gradual 7.6‰ shift. Collectively, the three sites display a trend of decreasing isotopic values (and therefore less C_4 prominence) from west to east for both the Late Pleistocene and early Holocene. Further, isotopic data from the three disparate locations suggest a Late Pleistocene environment dominated by C_3 plants, but to a lesser extent in western Kansas. The arboreal C_3 influence, as evidenced in

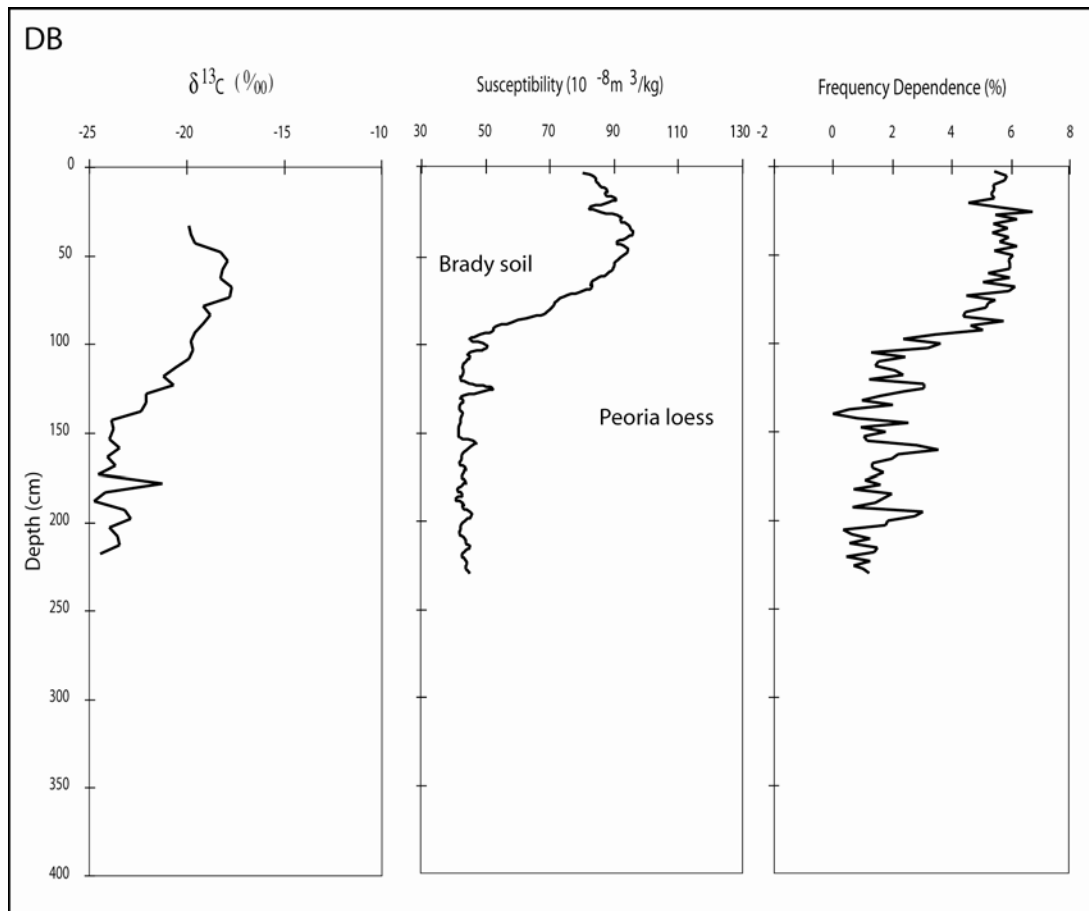


Figure 14. $\delta^{13}\text{C}$ from soil organic matter, susceptibility, and frequency dependence data for the DB site.

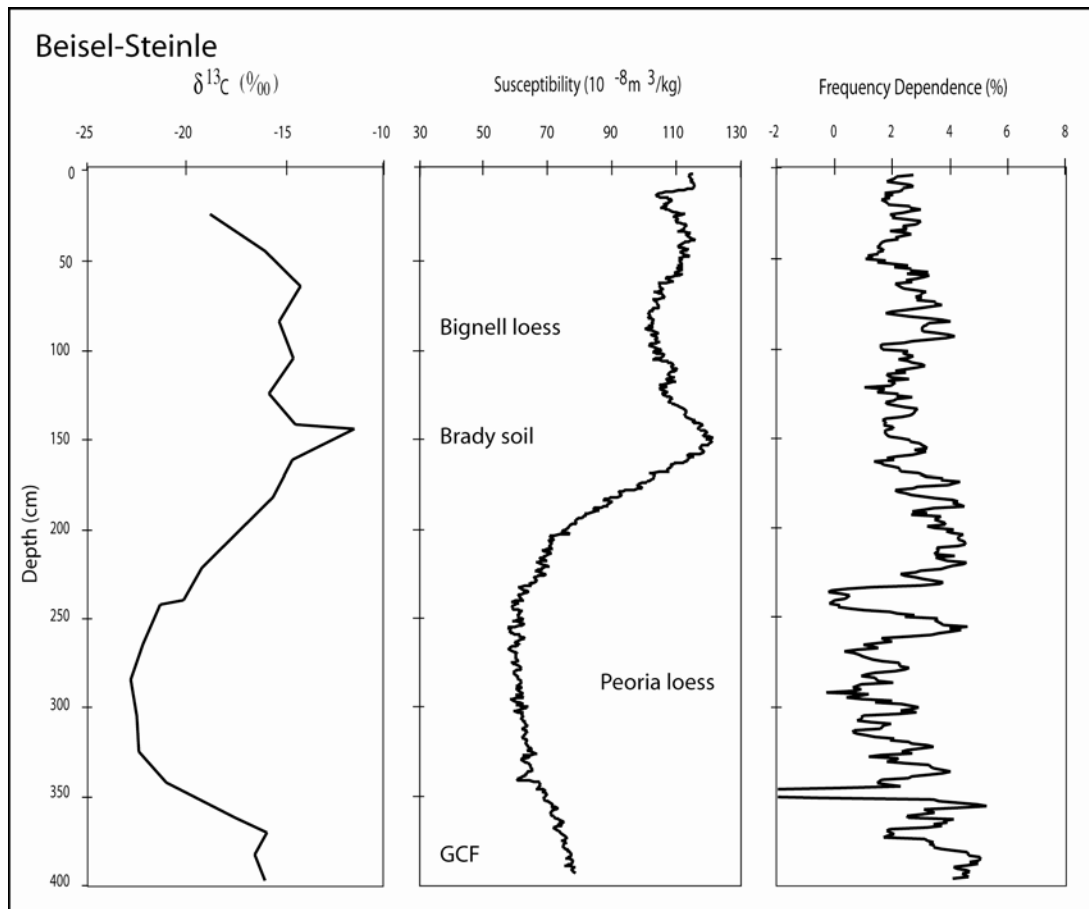


Figure 15. $\delta^{13}\text{C}$ from soil organic matter, susceptibility, and frequency dependence data for the Biesel-Steinle site.

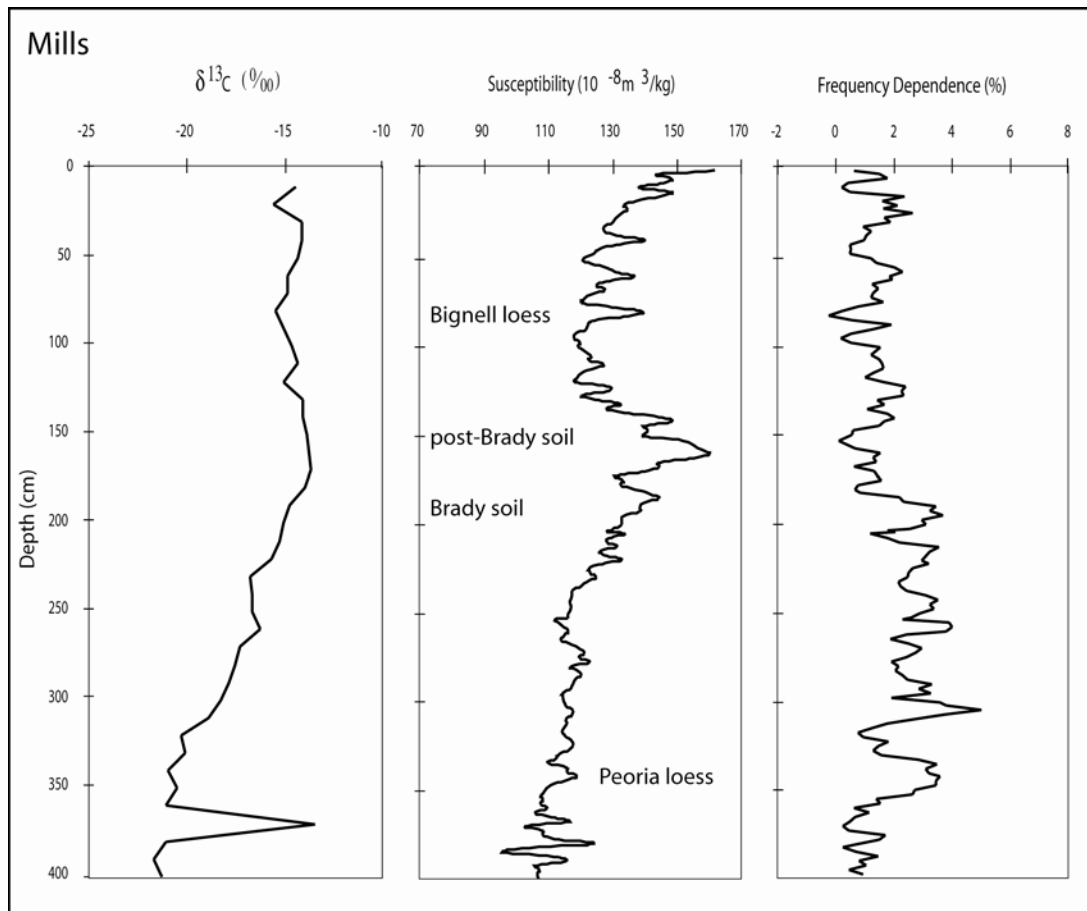


Figure 16. $\delta^{13}\text{C}$ from soil organic matter, susceptibility, and frequency dependence data for the Mills site.

the fossil pollen record from Muscotah Marsh in northeastern Kansas (Grüger, 1973), probably served to decrease isotopic values at DB, whereas the signal at Mills is likely a function of grassland dominated by C₃ types. As Brady pedogenesis came to an end in the early Holocene, vegetation had transformed to a C₄-dominated grassland in western Kansas and to a somewhat C₄-diluted C₃ plant community in eastern Kansas, certainly due, in part, to a diminished arboreal component, as evidenced at Muscotah Marsh (Grüger, 1973).

Growing season temperature is the primary climatic parameter explaining the distribution of grasses (Teeri and Stowe, 1978; Burke *et al.*, 1991; Nordt *et al.*, 2007), whereas the relative proportion of C₃ and C₄ grasses is represented by the $\delta^{13}\text{C}$ of soil organic matter (Fredlund and Tieszen, 1997b; Tieszen *et al.*, 1997). Using $\delta^{13}\text{C}$ derived from soil organic matter as a surrogate of the C₃-C₄ grass mix, Fredlund and Tieszen (1997b) made a comparison with July mean daily temperature (Figure 17). Using a regression of their data, the $\delta^{13}\text{C}$ values from east to west, DB, Beisel-Steinle and Mills yield Late Pleistocene July temperatures of 16, 18 and 20°C, respectively, and Holocene July temperatures of 24, 27, and 30°C. Although only approximations, these numbers indicate shifts in July mean daily temperatures on the order of 8-10°C for the three locations, values consistent with those derived by Fredlund and Tieszen (1997a). These estimates are slightly higher than Dutton's (1995) observation that groundwater temperatures of the High Plains aquifers during the Middle and Late Wisconsinian were 5 to 8°C cooler than those of the Holocene.

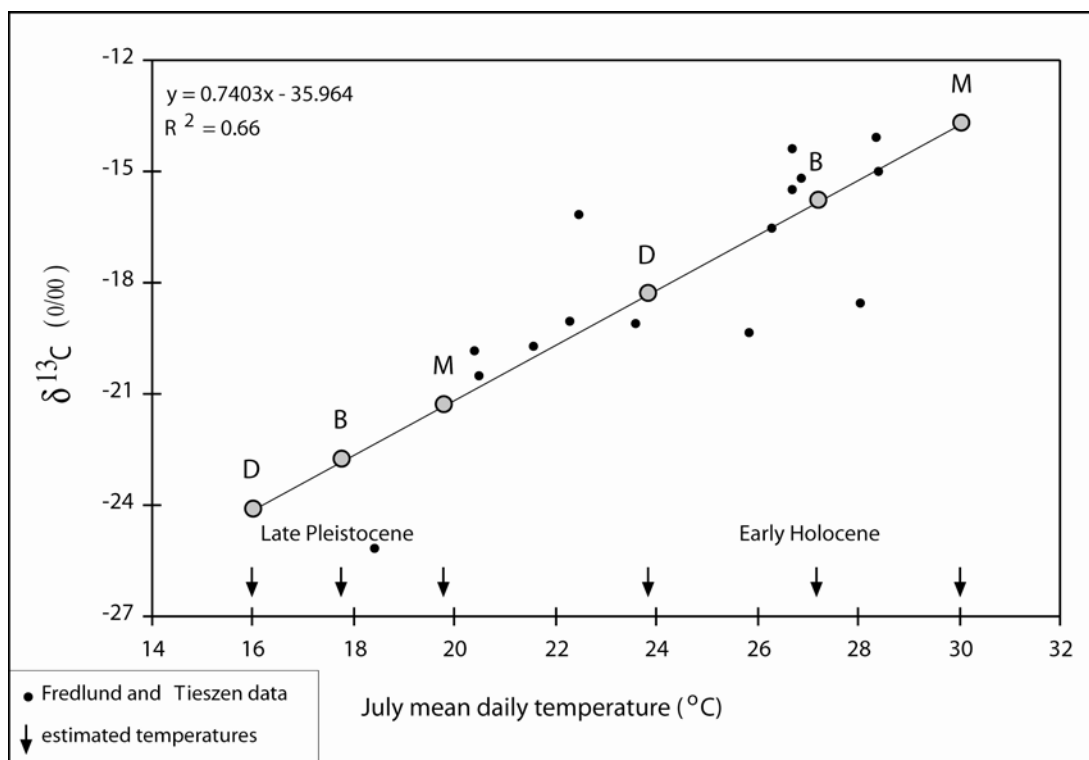


Figure 17. Correlation of July mean daily temperature with $\delta^{13}C$ from modern soil organic matter, and estimated Late Pleistocene and early Holocene temperature estimates. Letters D, B, and M along the regression line refer to the DB, Biesel-Steinle, and Mills sites, respectively.

Rock magnetic data

Analyses of magnetic susceptibility and frequency dependence for the DB, Beisel-Steinle, and Mills sites were conducted to characterize the Peoria Loess, Brady Soil, and Bignell loess. Both susceptibility and frequency dependence at DB are relatively low and uniform for the Peoria Loess, indicating a low concentration of magnetic minerals, including the ultrafine fraction, and a relatively high and constant accumulation rate (Figure 14). Pedogenesis triggered an increase in both parameters in the Brady Soil; welding of the surface soil to the Brady Soil has resulted in a plateauing of the parameters for the two soils. The small peaks within the Peoria Loess, which are present in both parameters, probably reflect a periodic decrease in accumulation rate that permitted an increased level of weathering occasionally expressed as incipient soil development. An alternative explanation could be differing amounts of magnetic minerals in loess source materials.

Susceptibility for Beisel-Steinle displays low values in the Peoria Loess, highest for the Brady Soil A horizon, somewhat lower in the Bignell loess, and a slight elevation for the surface soil (Figure 15). Frequency dependence is more variable, but exhibits generally low values for the Peoria Loess. The increases associated with the Brady and surface soils, displaced slightly below the corresponding susceptibility modes, indicate strong B horizon development through the accumulation of ultrafine (clay-size) magnetic materials. At Mills, susceptibility is lowest for the Peoria and Bignell loesses, with the Bignell loess being slightly higher and exhibiting four to five minor peaks that correspond to incipient soils (Figure 16).

Magnetic expression of the Brady Soil has been diminished by welding and overprinting from an early Holocene soil immediately above. A plateau in frequency dependence indicates that the two B horizons have fused, reducing the identity of the Brady A horizon and featuring the ubiquitous short-term drop in frequency dependence within the lower B horizon.

Mean susceptibility values derived for the three stratigraphic units at sites across the northern half of Kansas indicate differences in weathering and suggest an east-west trend (Figure 18). Susceptibility is lowest for Peoria Loess due to limited weathering and is highest in the Brady Soil. Intermediate values for the Bignell loess may reflect pre-weathering of the sediments, with loess derived from eroding Brady Soil surfaces being redeposited as loess. Because the distribution along valley margins suggests Bignell loess was transported, to some extent, out of stream valleys, this level of pre-weathering is not likely. Alternately, the slow deposition rate of the Bignell loess, relative to the Peoria Loess, permitted ample time for weathering and incipient soil development. Susceptibility of stratigraphic units appears to show increasing weathering intensity from east to west, however frequency dependence of susceptibility of the Brady soil suggests the opposite (Figure 18). Soil stratigraphy reveals that weathering is indeed greater in the east, as exhibited by the frequency dependence. Susceptibility of the Peoria Loess, then, is likely enhanced by magnetic mineralogy a source location to the west, the influence of which decreases to the east. The divergence of Brady and Bignell susceptibility from the Peoria susceptibility

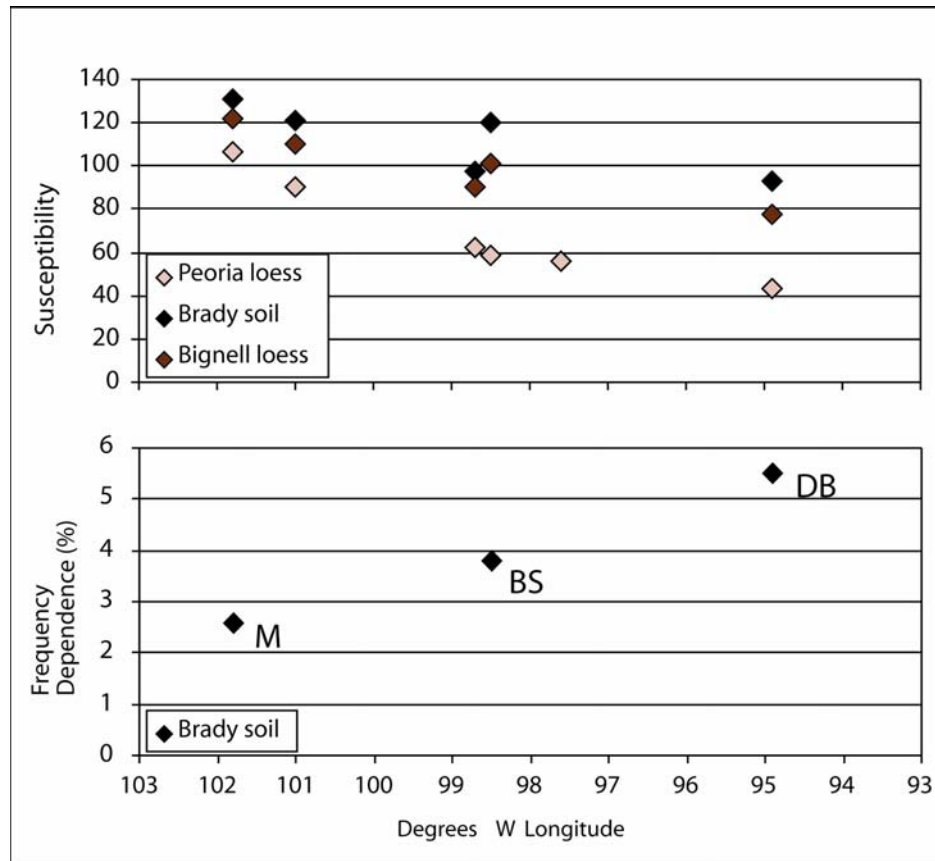


Figure 18. Average values of magnetic parameters by site and horizon with respect to longitude.

gives a better representation of the gradient of increasing pedogenic expression to the east.

Comparison of size-sensitive frequency dependence with concentration-sensitive susceptibility produces distinct clusters for the different lithostratigraphic and pedostratigraphic units. Biparametric representation of Beisel-Steinle magnetic data illustrates the stratigraphic segregation (Figure 19). Clustering of Peoria Loess samples reflects low and consistent susceptibility (c 59-65 $10^{-8} \text{m}^3/\text{kg}$), with frequency dependence covering a range from 0 to 4%. Because all samples were included, the upper end of frequency dependence involves samples transitioning into the underlying Gilman Canyon loess and the overlying Brady B horizon. Brady B horizon development is illustrated by the relatively high frequency dependence and progressively higher susceptibilities toward the A horizon. Overall, the highest group susceptibility characterize the Brady A horizon, with slightly reduced susceptibility common for the Bignell loess.

Summary and conclusions

GCMs simulate dramatic and rapid climate change in the central Great Plains during the Pleistocene-Holocene transition (COHMAP, 1988; Webb *et al.*, 1993). Warming, from as much as 10°C below mean annual temperatures of the present, transformed the C_3 -dominated grasslands and spruce parklands of the Late Pleistocene (Fredlund and Jaumann, 1987; Fredlund *et al.*, 1985; Hall and Valastro, 1995) into the C_4 -dominated grasslands of the Holocene (Küchler, 1964).

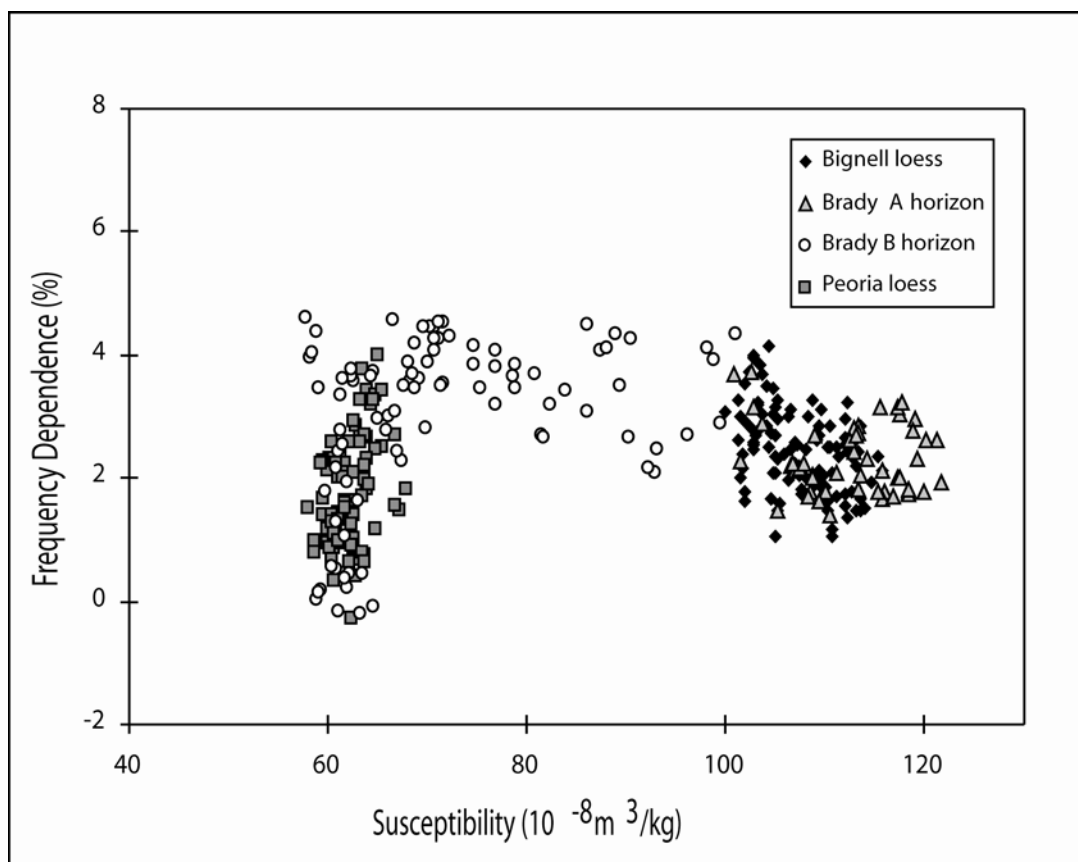


Figure 19. Biparametric representation of susceptibility and frequency dependence for the Peoria loss, Brady Soil A horizon, Brady Soil B horizon, and Bignell loess from the Biesel-Steinle site.

Widespread development of the Brady Soil from between about 13 ka and 9 ka implies that the loess-mantled uplands were stable during the climatic transition from the Pleistocene to the Holocene. Moreover, the entire landscape of the central Great Plains was apparently stable for this time period as indicated by the existence of a well-developed synchronous soil in alluvial fans (Johnson, 1999; Faulkner, 2002), terrace fills (Johnson and Martin, 1987; Johnson and Logan, 1990; Mandel, 1994; Johnson *et al.*, 1996), sand dunes and sheets (Arbogast and Johnson, 1998), and lunettes (Arbogast, 1996).

On a hemispheric scale, an abrupt decrease in atmospheric dust content in the Greenland ice core following the Younger Dryas climatic period implies reduced loess transport and deposition (Patterson and Hammer, 1987; Mason *et al.*, 2008), which correlates closely with the beginning of Brady pedogenesis. Meltwater pulses released in association with rapid deglaciation cooled the Gulf of Mexico between 15 ka and 11 ka (Flower and Kennett, 1990). This lowering of the Gulf's sea-surface temperatures may have decreased the intensity of storms in the North Atlantic, whereas the rising summer solar insolation would increase the temperature gradient between the Gulf of Mexico and the central United States; this condition could have resulted in increased monsoonal flow into the region (Kutzbach and Guetter, 1986; Maasch and Oglesby, 1990). Forman *et al.* (1995) speculated that the increased spring and summer precipitation on the High Plains would have resulted in a period of landscape stability and associated pedogenesis, as reflected in the aeolian stratigraphic record of the region. A paleoclimatic reconstruction from Moon Lake,

eastern North Dakota, indicates a slow transition from a fresh- to saline-water system between about 10 ka and 8.1 ka and a subsequent rapid and large change in salinity to 7.3 ka (Laird *et al.*, 1996); the period of slow transition suggests relative environmental stability. Similarly, the pollen record from Elk Lake, Minnesota, indicates relatively cool and moist conditions during this interval (Bartlein and Whitlock, 1993). The proxy record at Deep Lake, Minnesota, signals increased wind flow and associated aeolian deposition beginning about 8.1 ka (Hu *et al.*, 1999), which agrees with the record at Moon Lake. This change in climate likely brought about the end of landscape stability and associated Brady pedogenesis.

Stable carbon isotope and rock magnetic data from sites within the region exhibit a demonstrable response to climatic change during the Pleistocene-Holocene transition. Isotopic data show a rapid and steady increase from the upper Peoria Loess into the Brady Soil, where they assume near-Holocene levels. Future refinements in the application of stable carbon isotope analysis will aid in the extraction of a climatic record from loess (Philp and Johnson, 1999).

The two rock magnetic parameters applied display a strong response to Brady pedogenesis, and they differentiate between the “cold” climate loess of the Late Pleistocene and the “warm” climate loess of the Holocene. The yet unfulfilled potential of the rock magnetic approach to climatic reconstruction from loess is exemplified by recent research investigating the paleoclimatic significance of magnetite origins (Maher and Thompson, 1992), reconstructing the paleorainfall

record (Maher and Thompson, 1995) and estimating summer paleomonsoon intensity (Xiao *et al.*, 1995).

CHAPTER 4: Holocene Alluvial Cycles in the Loess Canyons of the Upper Republican River, Central High Plains

Abstract

Loess stratigraphy of the central Great Plains offers a terrestrial cognate to the deep sea and ice cores for deducing environmental change in the Late Pleistocene and Holocene. Resolution of the loess record, however, decreases during the Holocene in many locations with the slowing of the loess flux. Fluvial morphology, sand dune mobilization, and other landscape settings can more precisely identify periods of Holocene environmental change. Loess canyons draining to the upper Republican River in northwestern Kansas and adjacent Nebraska have formed by gully cut-and-fill cycles during the Holocene. Massive erosion occurred during the Altithermal and ended between 5 and 4 ka with the aggradation of the T-2 sediment package preserved as terrace remnants throughout the canyons. The lower T-2 soil was dated between 3.2 and 2.9 ka, indicating a period of valley stability that was followed by alluviation. Upper T-2 fill is punctuated with periods of soil formation that are not consistent among all sites, but show concentrations around the dates of 1.8, 1.5, and 1.2 ka. Thin deposits of undifferentiated alluvium cap the T-2 which was dissected by gully erosion around 1.0 ka corresponding to the Medieval Warm Period. The resulting T-1 shows little pedogenic alteration and is also cut by a gully of presumed historic origin.

Introduction

Contrasting sharply with the gently undulating loess topography of the High Plains, spectacular gully erosion in extreme northwestern Kansas and adjacent parts of Colorado and Nebraska has earned the regional name, “the Breaks.” Located at the fringe of the loess plateau, the Breaks form a maze of deep canyons draining into the shallow Republican River along the Nebraska state line. In northern Cheyenne County, Kansas, episodes of gully advance have carved out canyons over 100 m deep that are expanding into the loess-mantled upland. The record preserved in the canyon alluvial sediments suggests that in recent geologic history, periods of landscape stability and soil formation have been abruptly ended by widespread episodes of intense erosion. Stratigraphy lacking buried soils indicates rapid aggradation due to headward gully extension or erosion from exposed canyon sidewalls. Terrace remnants near the canyon walls provide a record of prior fills eroded by gully widening.

Previous studies documented synchronous alluvial responses (Knox, 1983; Johnson and Martin, 1987; Johnson and Logan, 1990; Mandel, 1992a, b) and suggested regional climatic changes as the major contributing factor (Davis, 1902; Schumm and Lichty, 1963). In an area where traditional records of Holocene climatic history (*e.g.* ice cores, lake varves, speleothems) are sparse, or, in the case of loess stratigraphy, of coarse resolution, fluvial systems can provide records of landscape change on the order of centuries. According to Knox (1993), climate changes on the order of 1-2°C, as occurred during the Holocene, can affect vast changes in the flood

regimes of fluvial systems and in stream morphology. Climatic inferences derived from nearby aeolian deposits and regional alluvial systems can help provide the link between activity in the loess canyons and drought.

The objectives of this investigation are to (1) document periods of aggradation, stability, and gully growth in the canyons of the Breaks, using Hay Canyon as a model, (2) compare the activity in Hay Canyon to that in surrounding canyons and other landscape positions nearby, and (3) correlate these findings with other regional environmental records from the Holocene.

Study area

The Breaks lie in the High Plains of the central United States (Schoewe, 1949) and dominate the northwestern corner of Cheyenne County, Kansas, including small portions of Dundy County, Nebraska and Yuma County, Colorado (Figure 20). Much of this region consists of a broad, slightly undulating loess plateau rising to 1,160 m above sea level with local relief of 3 to 15 m. Shallow depressions holding ephemeral “playa” lakes account for much of this relief (Johnson *et al.*, 2009). Numerous tributaries dissect the interior loess tableland and join the major drainages of Cheyenne County. The Arikaree River, which rises in eastern Colorado, cuts through the extreme northwestern part of the county and joins the North Fork Republican River, creating the Republican River near Haigler, Nebraska (Figure 21). Immediately north of the state line near Benkelman, Nebraska, the South Fork joins the Republican River and flows eastward (Prescott, 1953).

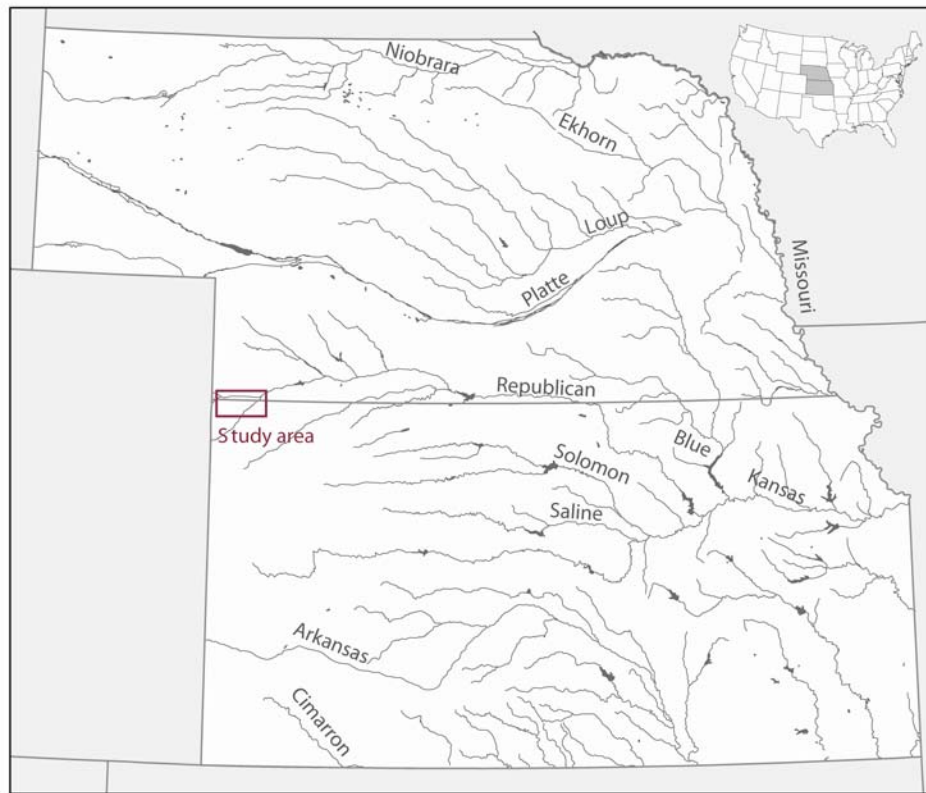


Figure 20. The study area location within the upper Republican River basin.

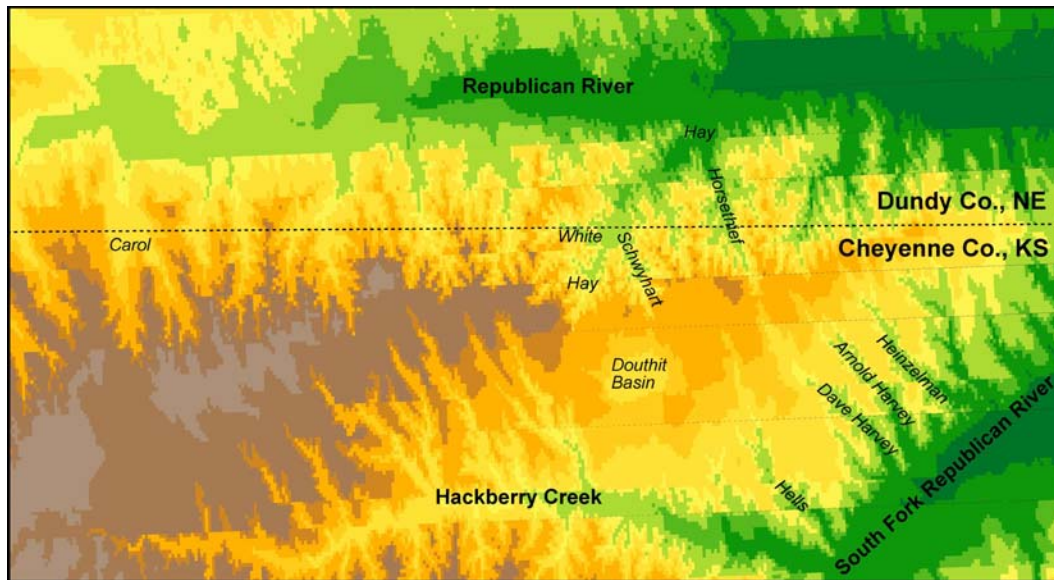


Figure 21. Digital Elevation Model (DEM) showing the location of Hay Canyon in the Breaks region, which extends along the south side of the Republican River. The south fork canyons (Hells, Dave Harvey, Arnold Harvey, and Heinzelman) drain southeast into the South Fork Republican River between Hackberry Creek and the Nebraska state line.

The rugged, deeply dissected uplands along the northern part of the county, known as the Breaks, are characterized by dendritic, north-draining canyons of the Arikaree and Republican Rivers. These canyons and the more subtle, linear canyons draining into the South Fork Republican River are sculpted through multiple episodes of gully cut and fill (Figure 21). Many of the box canyons in the Breaks express 35 m of local relief or 100 m between the height of the loess plateau and the mouth of the canyon up to 10 km away (Figure 22). Individual gully headwalls within the level canyon valleys are commonly 3-5 m high, but heights of these features increase in some canyons to 12-15 m near the intersection with the uplands. Active gullies are currently dissecting relatively flat valley floors (T-1) ranging in width from 3 m near the head to 100 m at their intersection with major drainages. Terrace surfaces (T-2) preserve remnants of older, wider valley floors, indicating multiple episodes of Holocene cut and fill in the canyons.

Hay Canyon, a prominent drainage system in the Breaks, drains approximately 40 km² of northern Cheyenne County, Kansas, into the Republican River in Dundy County, Nebraska. The main branch, and a nearly-equivalent tributary, Schwyhart Canyon, drain north where they are joined by White Canyon from the west. There the main valley turns northeast before receiving a tributary, Horsethief Canyon, from the south. From that confluence the valley flattens out and enters the Republican River valley.

The flat valley floor (T-1) of lower Hay Canyon is cut by an actively-advancing (T-0/T-1) gully, producing a new valley floor (T-0) connecting to the



Figure 22. View of the upper Hay canyon system, north toward the Republican River. Note active (T-1/T-2) gullies advancing through T-2 fill toward the upland.

Republican River. The T-1 surface stretches nearly unbroken into the upper reaches of the canyon where remnants of an older, T-2 surface cling to the sides of the canyon and near canyon confluences. The upper (T-1/T-2) gully head cut is well-advanced into the upland. The terminal ends of gullies in Hay and Schwyhart Canyons are “stepped” into distinct upper and lower headwalls, likely due to the resistance from encountering Pierre Shale at that depth.

South Fork canyons, in contrast, tend to be narrower and nearly parallel, draining southeast into the South Fork Republican River. Overall, headwalls are lower, and the steeper main valleys account for more of the overall relief. As in the Breaks, the T-0/T-1 gully has breached only the lower quarter of most canyons; however, the T-1/T-2 gullies fall short of the extent found in Hay Canyon.

Regional Geology

In some locations, erosion of the canyons through the loess cover has exposed the Neogene Ogallala Group and the underlying Cretaceous Pierre Shale. The Ogallala in Cheyenne County is highly variable, porous sandstone made up of calcite-cemented quartz and feldspar grains. Buried Pierre Shale forms an impermeable layer under the porous Ogallala and dictates the location of the regional water table. The upper surface of the shale in the study region dips gently to the south, generating springs in the South Fork canyons that are absent in the Breaks.

Quaternary loess, blanketing much of the region’s uplands, contains five major stratigraphic units: undifferentiated Pre-Illinoian loess, Illinoian Loveland

loess, including the interglacial Sangamon Soil, the Middle Wisconsinan Gilman Canyon Formation loess and soils, Late Wisconsinan Peoria Loess and overlying Brady Soil, and the Holocene Bignell Loess. Two major sand sheets lie to the north (Nebraska Sand Hills) and northwest (in Colorado), and have contributed to the Quaternary dust flux during times of dune mobilization (Mason *et al.*, 2003; Miao *et al.*, 2007b).

Climate

The weather station at Saint Francis, the county seat of Cheyenne County, receives 46.6 cm of precipitation in an average year (High Plains Regional Climate Center, 2002). As is characteristic of the semi-arid High Plains though, actual precipitation amounts can vary drastically from the norm; the last 100 years of annual precipitation data show a standard deviation from the mean of 12.3 cm. Summer rainfall occurs mainly associated with infrequent, but intense cyclonic storms. Snowfall in the winter months accounts for a small amount of overall yearly precipitation, but is vital in preventing winter winds from further desiccating the soil. Spring snow melt helps to replenish lost soil moisture. Average maximum temperatures peak in July and August at 33.4°C and 32.3°C (High Plains Regional Climate Center, 2002). High temperatures and low relative humidity result in frequent periods during the growing season when soil moisture fails to meet the evapotranspiration needs for crops.

Methods

Stratigraphic Investigations

Six sites from the T-2 terrace sediments of Hay Canyon were selected for documentation using exposed faces of the canyon walls: Cave, Upper Cave, West Hay, Scott, Fred, and Mills 3. Cores were extracted from the T-2 and T-1 alluvium at the Cave site, adjacent to the studied profile. Two exposures of T-1 sediment were sampled in lower Hay Canyon at the Williams site: one where the new gully eroded close to the valley wall, and one from a remnant butte of T-1 sediment left within the new gully channel. Three T-2 sites were described in the South Fork region: Lower Hells, Lower Dave, and Lower Arnold. Cores were also taken from the T-2 and T-1 sediments at the Lower Hells site.

Surveying

Using a traditional transit and stadia rod technique, relief within the study canyons was documented along the longitudinal profiles. The length of Hay Canyon was surveyed from the Republican River to the headwall along its main axis. Surveys of relevant side canyons were then tied to that of the main axis for the analysis of valley gradients and site locations within the system. In the upper reaches of the canyons, where terrace fragments occurred with regularity, it was possible to include surveys of the T-2 surfaces and the paleovalley gradients they represent. Transect initiation points were determined by GPS, but further use of satellites was often limited by the steep canyon walls. The lowermost 2.5 km of Hay Canyon was

determined from the 1961 USGS topographic map, in order to document the historic progress of the T-0/T-1 gully headwall.

Numerical age control

Conventional radiocarbon dating of soil humates was used for most of the absolute time control at the study localities. Samples were collected from 5 cm vertical intervals encompassing the uppermost portion of Ab horizons, or the horizon showing the strongest pedogenic signature if a soil complex exhibited multiple Ab horizons. Samples were pre-treated in the University of Kansas Soil Analysis Laboratory by the methods described in Johnson and Valastro (1994) and then submitted to the Illinois State Geological Survey Radiocarbon Laboratory for dating and correction for isotopic fractionation of carbon. Calibration from radiocarbon years to calendar years was calculated using the calibration curve of Fairbanks *et al.* (2005).

Supplemental time control was provided by optically stimulated luminescence (OSL) dating as an alternative to a carbon-based signal. Huntley *et al.* (1985) developed the technique as an extension of the theories behind thermo-luminescence (TL) dating. Exposure of mineral grains to light (*e.g.* during sediment transport) clears the electrons from radiation charge-trapping sites within the crystal lattice of quartz and feldspars. After the sediment becomes buried and is no longer exposed to light, natural environmental radiation from uranium (U), thorium (Th) and potassium (K) deposits charges back into the quartz grains (Teeuw *et al.*, 1999). By measuring

the background rate of radiation from the sediment and the total concentration within the quartz grains, the time elapsed since burial of the sediment can be calculated by the formula from Eriksson *et al.* (2000):

$$\text{Burial time (years)} = \text{Burial dose (Gy)} / \text{Dose rate (Gy/year)}$$

If insufficient bleaching (exposure to light) occurs prior to deposition and burial of sediment, the sample may overstate the time since burial by reporting a higher burial dose than the time of burial would naturally allow. Because the distance of transportation in the Breaks is short between erosion in the uplands and deposition as alluvium in the canyon bottoms, insufficient bleaching may be a problem in this study site. Three OSL samples were taken from a site with good radiocarbon time control to test the application of this method to the region and further validate the radiocarbon ages used in the study.

Canyon morphology and stratigraphy in the Breaks

Hay Canyon system

The upper end of Hay Canyon, 9.1 km distant from the river, is dominated by a vertical gully head wall that drops abruptly 13.5 m, revealing Quaternary loess stratigraphy (Figure 23). Only 30 m down-canyon, the valley floor drops 11 m again into a second, more gradual gully headwall (the T-1/T-2 gully). This gully's

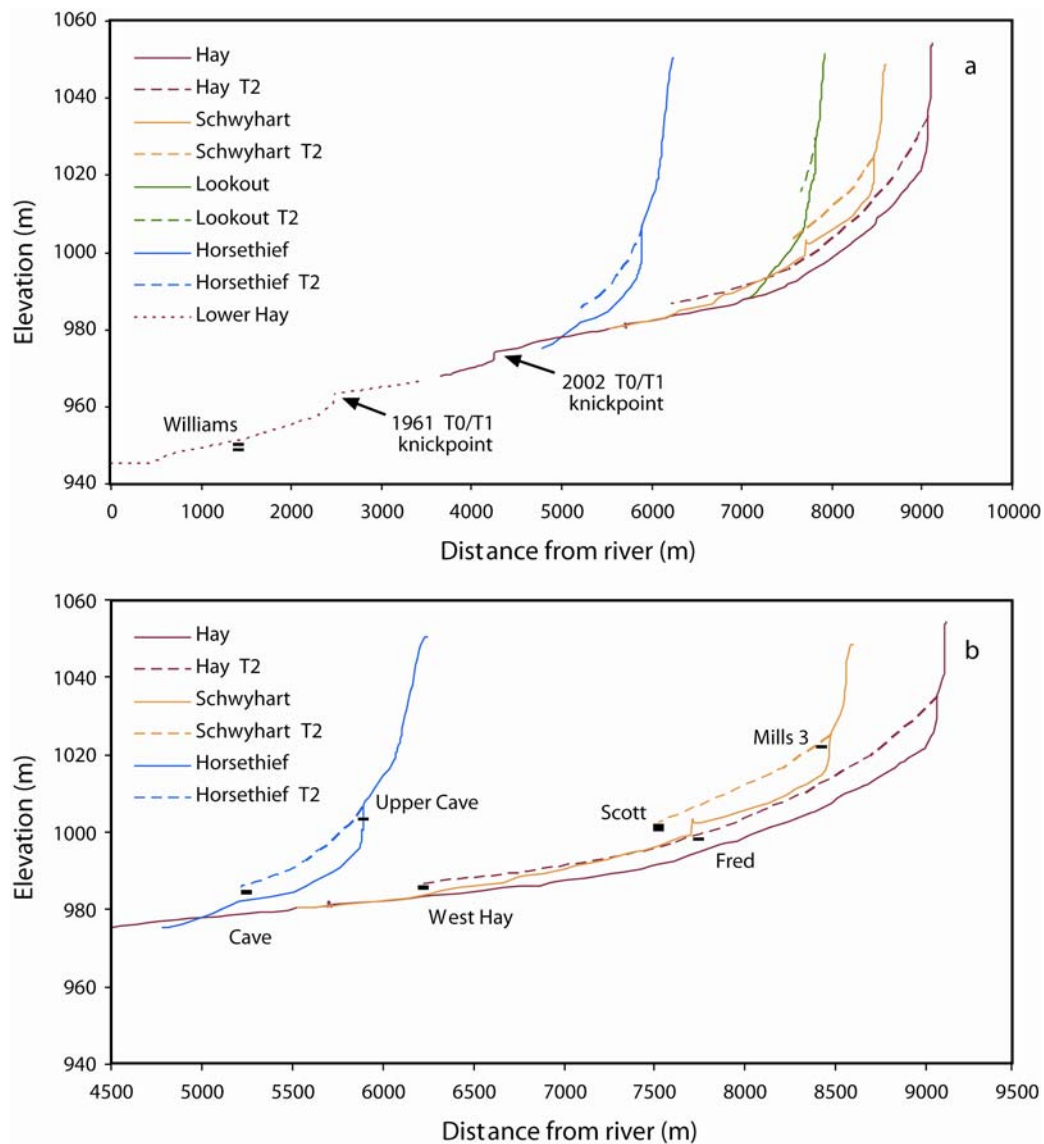


Figure 23. a) Surveys within Hay Canyon (at 40x vertical exaggeration) show the relative progress of the T-1/T-2 gully toward the upland as well as preserved T-2 terrace remnants in the upper reaches of each branch. Surveys also reveal 1.7 km of migration for the T-0/T-1 knickpoint from the 1961 USGS topographic map to 2002. b) Locations of each sample site within the Hay Canyon tributaries.

progress has been hampered by encountering the more resistant Pierre Shale that underlies the loess above Hay Canyon. Small remnants of a T-2 terrace surface sit 10 m above the current valley floor. Further down-canyon (7.8 km from the river) where the T-2 terrace height lowers to 4.5 m above the valley floor, the Fred site exhibits a well-developed buried soil in a long narrow finger of T-2 alluvium, separating the main branch of Hay Canyon from a tributary entering from the west. A radiocarbon sample collected from the upper part of this soil at 1.3 m (Figure 24) yielded an age of $1,860 \pm 80$ cal yr BP (Table 4). Between the Fred site and the T-1/T-2 gully, the average slopes of the T-2 and T-1 surfaces are 2.7 and 1.2%, respectively. The West Hay site features another preserved T-2 remnant found in White Canyon immediately above its confluence with Hay Canyon (6.2 km from the river). At a depth of 1.51 m, a soil yielded a radiocarbon age of $1,590 \pm 80$ cal yr BP, whereas a sample of charcoal above the soil (1.24 m) yielded a younger age of $1,480 \pm 80$ cal yr BP (Figure 23).

Entering Hay Canyon above the Fred site are a series of short, steep side canyons with well-preserved T-2 terraces in their upper reaches. One of these, Lookout Canyon, was selected for surveying (Figure 25). The total length of Lookout Canyon from the upland to its confluence with Hay Canyon is 0.9 km (Figure 23); consequently, both the current (T-1) and past (T-2) valley floors are steep compared with longer, more substantial tributaries within Hay Canyon. The T-2 surface between the T-1/T-2 gully headwall and the lower end of the preserved terrace slopes at 7.7%. While the T-1 valley floor immediately below the measured T-2 was broken

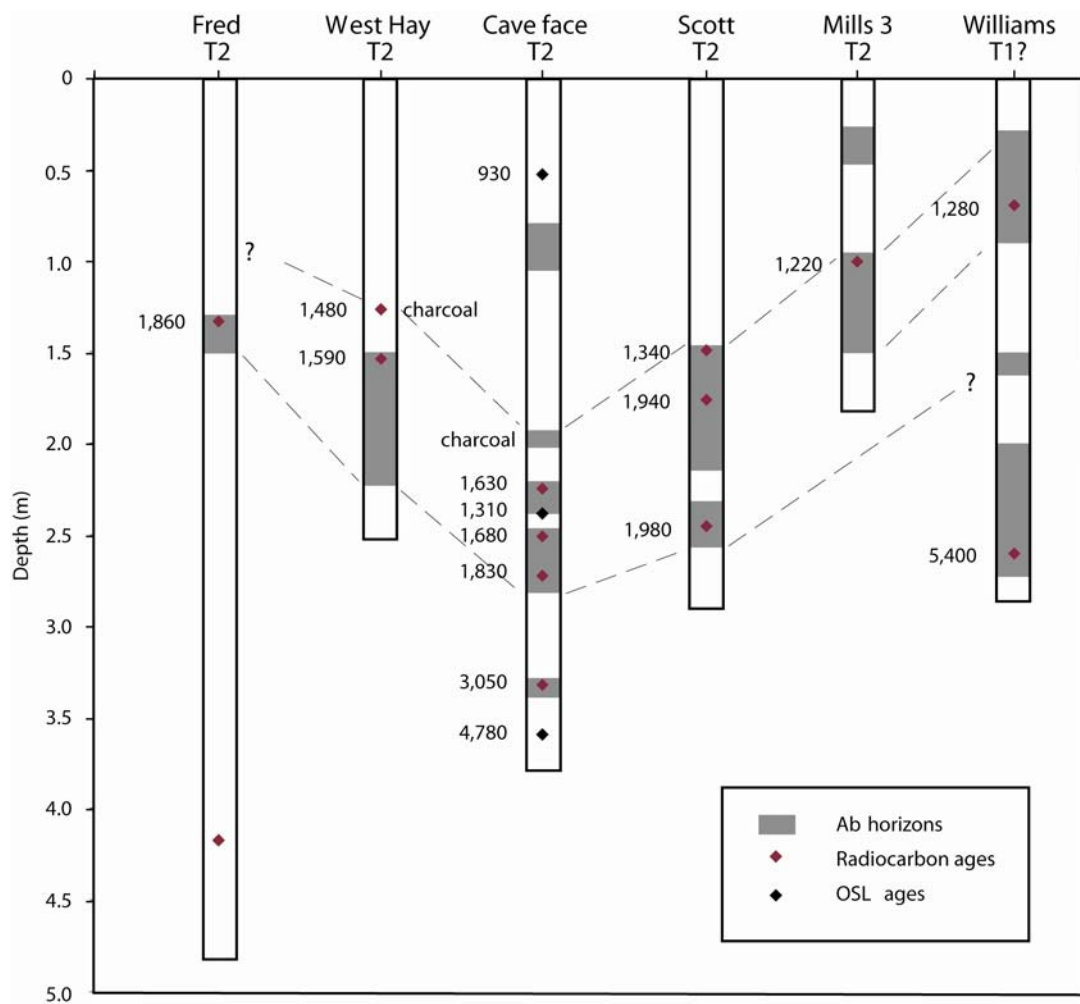


Figure 24. Correlations between stratigraphy of each Hay Canyon site. See Tables 5 and 6 for ages.

Table 4. Radiocarbon ages from Cheyenne County, KS and Dundy County, NE

Sample Number	Depth (cm)	Laboratory Number	$\delta^{13}\text{C}$ (‰)	Corrected age a (yr BP)	Std. dev.	Calendar Age b (yr BP)	Std. dev.
Republican River Breaks							
Cave 222	222	ISGS-5241	-17.0	1,720	70	1,630	90
Cave 248	248	ISGS-5250	-16.4	1,760	70	1,680	90
Cave 269	269	ISGS-5238	-15.2	1,890	70	1,830	80
Cave 329	329	ISGS-5249	-18.5	2,910	70	3,050	110
U. Cave 368	368	ISGS-5466	-17.2	1,630	70	1,520	80
Fred 130	130	ISGS-5474	-17.2	1,920	70	1,860	80
Mills 3-97	97	ISGS-5473	-16.5	1,290	70	1,220	70
Scott 146	146	ISGS-5489	-17	1,440	70	1,340	50
Scott 173	173	ISGS-5494	-16.6	1,990	70	1,940	80
Scott 242	242	ISGS-5465	-16.1	2,030	70	1,980	90
West Hay 124	124	ISGS-5479	-17.4	1,590	70	1,480	80
West Hay 151	151	ISGS-5472	-16.6	1,680	70	1,590	80
Williams 67	67	ISGS-5482	-15.2	1,360	70	1,280	50
Williams 257	257	ISGS-5480	-15.2	4,670	70	5,400	100
Workman 152	152	ISGS-4683	-15.5	1,610	70	1,500	80
South Fork Republican River Canyons							
L. Arnold 106	106	ISGS-5497	-18.8	1,670	70	1,570	80
L. Dave 102	102	ISGS-5499	-15.2	1,240	70	1,170	90
L. Hells T2 45	45	ISGS-5498	-15.8	1,180	70	1,100	90
L. Hells T2 81	81	ISGS-5495	-16.6	1,860	70	1,800	80
L. Hells T2 685	685	ISGS-5496	-17.5	3,190	70	3,410	70
Other Cheyenne County, Kansas Sites							
Douthit B 89	89	ISGS-5490	-18.3	1,680	70	1,580	80
Douthit B 260	260	ISGS-5491	-18.2	10,010	140	11,520	270
Douthit B 271	271	ISGS-5492	-20.1	9,840	110	11,250	120
Douthit B 281	281	ISGS-5493	-21.6	9,480	130	10,760	220
Landfill 183	183	ISGS-5099	-17.6	9,960	70	11,380	140
Loop 105	105	ISGS-5100	-14.5	1,200	70	1,120	90

a Ages corrected for isotopic fractionation.

b From Fairbanks *et al* 2005

Table 4. continued

Sample Number	Depth (cm)	Laboratory Number	$\delta^{13}\text{C}$ (‰)	Corrected age a (yr BP)	Std. dev.	Calendric Age b (yr BP)	Std. dev.
UPLANDS							
Jones 310	310	ISGS-5263	-17.1	7,480	70	8,310	70
Jones 385	385	ISGS-5266	-18.1	10,820	70	12,730	60
KWP 113	113	ISGS-5477	-15.6	7,940	70	8,780	140
KWP 173	173	ISGS-5254	-17.9	9,960	70	11,380	140
Miller 175	175	ISGS-4390	-19.4	14,090	70	16,470	150
Miller 265	265	ISGS-4392	-17.3	23,320	270	27,970	330
Miller 335	335	ISGS-4391	-19.6	24,830	160	29,760	280
Mills 150	150	ISGS-3922	-14.0	3,970	70	4,430	90
Mills 175	175	ISGS-3926	-13.6	4,940	70	5,670	70
Mills 225c	225	ISGS-3929	-16.5	7,940	70	8,780	140
Mills 263	263	ISGS-3917	-17.1	9,390	70	10,610	90
Mills 282c	282	ISGS-3915	-17.5	9,720	70	11,150	90
State 203	203	ISGS-5469	-15.8	1,570	70	1,450	80
State 365	365	ISGS-5470	-15.1	5,590	70	6,370	70

c Johnson and Willey, 2000.

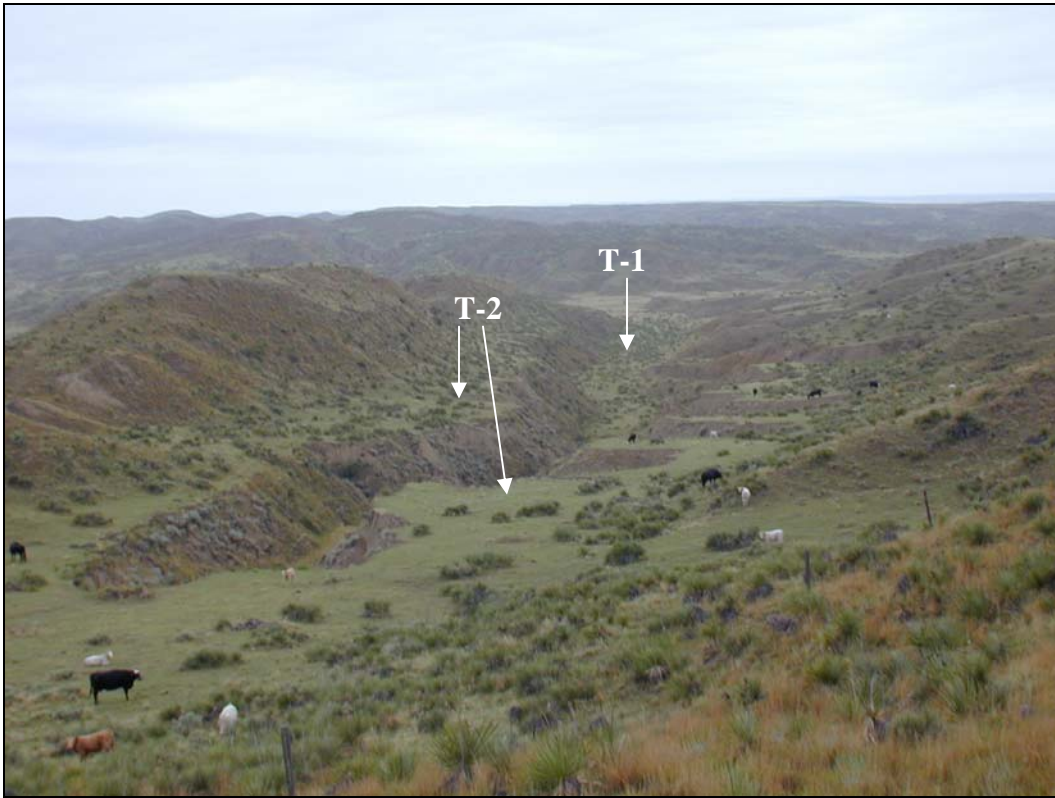


Figure 25. Lookout Canyon exhibits well-preserved T-2 terrace remnants, which yield a paleovalley gradient of 7.7%, in contrast to 2.9% for the present valley floor.

into step gullies as a result of encountering Pierre shale, the segment of undisturbed T-1 immediately below the shale sloped at 2.9%.

Schwyhart Canyon, the eastern branch of upper Hay Canyon, has a similarly steep T-1/T-2 gully headwall at its upper end and fragments of T-2 terraces present along the upper 1.5 km (Figure 23). The slope of the valley floor represented by the T-2 is 2.4% compared with the current (T-1) valley floor slope of 1.2%. The Scott site was excavated from T-2 terrace remnant 7.5 km up the canyon from the Republican River. The terrace stands 4 m above the T-1 surface, but only the upper 2.5 m were described and sampled because colluvium had buried the base of the terrace. The upper 1.5 m was made up of distinct bands of sediment showing no sign of pedogenic alteration, whereas a very weakly developed Ab1 yielded an age of $1,340 \pm 50$ cal yr BP immediately below which was a better defined Ab2 which dated to $1,940 \pm 80$ cal yr BP and a Ab3 dating to $1,980 \pm 90$ cal yr BP (Figure 24). The Mills 3 site is a small T-2 remnant perched high along a side tributary to Schwyhart Canyon above the Scott site. A soil at 0.27 m, covered by overburden from the canyon sidewall, is presumed to be a near-modern surface, but below that, a soil buried at 0.95 m dated to $1,220 \pm 70$ cal yr BP.

Both T-2 and T-1 stratigraphy were described at the Cave site, 5.3 km from the Republican River in Horsethief Canyon (Figure 26). An exposed T-2 face at the confluence of two equal branches of Horsethief Canyon was sampled for both radiocarbon and OSL dating. An OSL sample from unaltered alluvium at 0.5 m dated to 930 ± 60 cal yr BP (Figures 24, 27; Table 5). Two welded incipient soils below that



Figure 26. Sampled profile and T-2 core locations for the Cave site. A T-1 core taken from the valley bottom in the foreground yielded pedogenically unaltered alluvium with lenses of Ogallala Formation and Pierre shale eroded from the T-1/T-2 gully headwall.

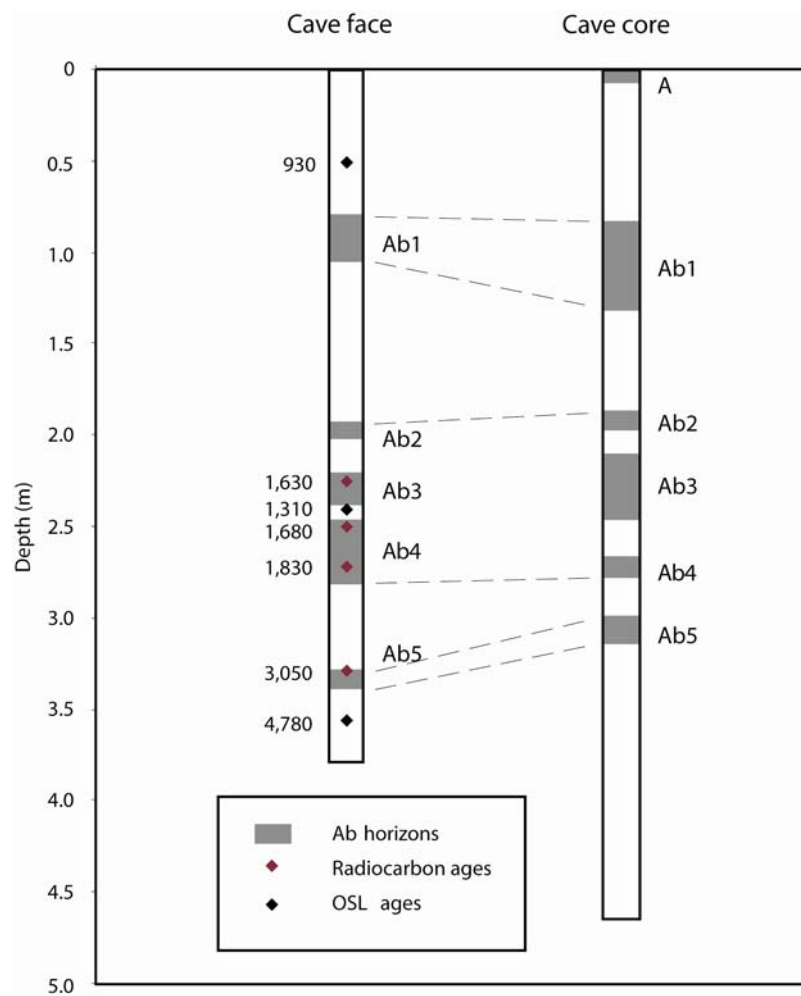


Figure 27. Stratigraphic correlation between the Cave face and core.

Table 5. Optically-stimulated luminescence (OSL) ages from Cheyenne County, Kansas

Sample Number	H ₂ O (%) [*]	K ₂ O (%)	U (ppm)	Th (ppm)	Cosmic (Gy)	Dose Rate (Gy/ka)	D _e (Gy)	Age
Alluvium								
Cave 50	2.9	2.45	3.1	12.2	0.24	3.69±0.10	3.43±0.1	930±60
Cave 234	2.9	2.55	3.2	13.0	0.18	3.80±0.11	4.99±0.1	1310±90
Cave 356	5.1	2.51	2.8	11.6	0.16	3.46±0.10	16.55±0.8	4780±370

were undated, but a cluster of three lower soils yielded radiocarbon dates of $1,630 \pm 90$ cal yr BP at 2.22 m, $1,680 \pm 90$ cal yr BP at 2.48 m, $1,830 \pm 80$ cal yr BP at 2.69 m, and in the lowermost soil, an age of $3,050 \pm 110$ cal yr BP was derived. An OSL age of $1,310 \pm 90$ cal yr BP at 2.34 m was 350 years young relative to the radiocarbon ages bounding the sample. The pedogenically young sediments below the bottom soil yielded an OSL age of $4,780 \pm 370$ cal yr BP at 3.56 m. A core collected through the same T-2 terrace penetrated an additional 1 m below the lower-most age without reaching additional buried soils (Figure 27). A second core taken in the adjacent T-1 sediments yielded 2.08 m of silty alluvium intercalated with layers of sandy alluvium derived from the Ogallala Formation exposed at the knickpoint of the T-1/T-2 gully. Further up Horsethief Canyon, 0.7 km above the Cave site, the T-2 surface steepens as it approaches the uplands. At this Upper Cave site, two buried soils, separated by more than 2 m of alluvium, were visible in the gully headwall. The lower soil dated to $1,520 \pm 80$ cal yr BP at 3.68 m and, though more deeply buried than at the Cave site, likely represents a welded expression of the three-soil-complex of a similar age at the Cave site.

The Williams site in the lower reaches of Hay Canyon (1.4 km from the river) has been eroded by the progression of the T-0/T-1 gully. Not only has the gully extended past the Williams site, but it has also eroded laterally, encompassing nearly the entire width of the valley bottom. An exposed face on the east gully wall had 0.3 m of pedogenically young fill capping a soil (Ab1) dating to $1,280 \pm 50$ cal yr BP (Figure 24). An age of $5,400 \pm 100$ cal yr BP from a depth of 2.57 m represents the

lower-most of a three soil complex. Immediately across from this profile stands an isolated T-1 remnant within the gully channel. This sediment package shows characteristics of recent deposition with layers of pedologically unaltered alluvium. The contrast between the expected and recognizably young T-1 within the channel and the characteristic T-2 ages of the soils within the Williams face imply that the sampled east wall of the gully has actually breached older sediment that has melded with the height of the T-1 in the lower reaches of the canyon.

The T-0/T-1 gully in lower Hay Canyon does not appear to have affected valley slopes significantly. Below that gully knickpoint both the T-0 and remaining T-1 surfaces had valley slopes of 0.5%. In the upper reaches of Hay, Schwyhart, Lookout, and Horsethief Canyons, however the T-1/T-2 gully has lowered the valley floor gradient between 0.6 and 4.8%. The minor tributary of Lookout Canyon accounts for the greater change, while Hay, Schwyhart, and Horsethief Canyons fall within a smaller range of 1.2 to 0.6%. The differential slopes between the T-1 and T-2 could reasonably account for the merging of the two surfaces in lower Hay Canyon as suggested for the Williams site.

Hay Canyon chronology

No alluvial sediments in the Breaks were found to record ages older than 5.4 ka. An erosional phase must pre-date the development of the T-2, but it is unknown whether this erosion accounted for canyon initiation, or if any previous cut and fill cycles had occurred. No evidence of a T-3 was revealed in this investigation.

The 5.4 ka age from the lower soil at the Williams site could represent the initiation of T-2 alluviation, but at least before 4.8 ka the T-2 had begun accumulating at the Cave site with a period of soil formation occurring at 3.0 ka (Figure 28). Additional coring into the T-2 remnants would establish the distribution of this soil in the canyon system, but current methods did not investigate deep enough to establish its extent. Following 3.0 ka, alluviation resumed, slowing to allow the development of a well-documented complex of soils dating between 1.8 and 1.2 ka. Subsequently, the T-1/T-2 gully progressed through much of the canyon bottom, establishing the current T-1 as the new alluvial surface. Following 1.2 ka, the T-1/T-2 gully has progressed 9.1 km up the main axis of Hay Canyon and encroached upon the uplands at the head of the canyon. The resulting T-1 alluvium is less than 1,000 years old and shows little pedogenic alteration.

The 2002 survey of the T-0/T-1 gully head indicated progress of 4.2 km into the Hay Canyon valley. By comparison, the 1961 USGS topographic map showed the location of the same gully head 2.5 km from the river. In 41 years the gully had moved 1.7 km with a rate averaging about 40 m/yr. Shallow exploration within the T-0 alluvium reveals the recent character of that fill. Well-defined sedimentary structures and buried car parts confirm that deposition of the T-0 corresponds to the historic erosion observed for the T-0/T-1 gully.

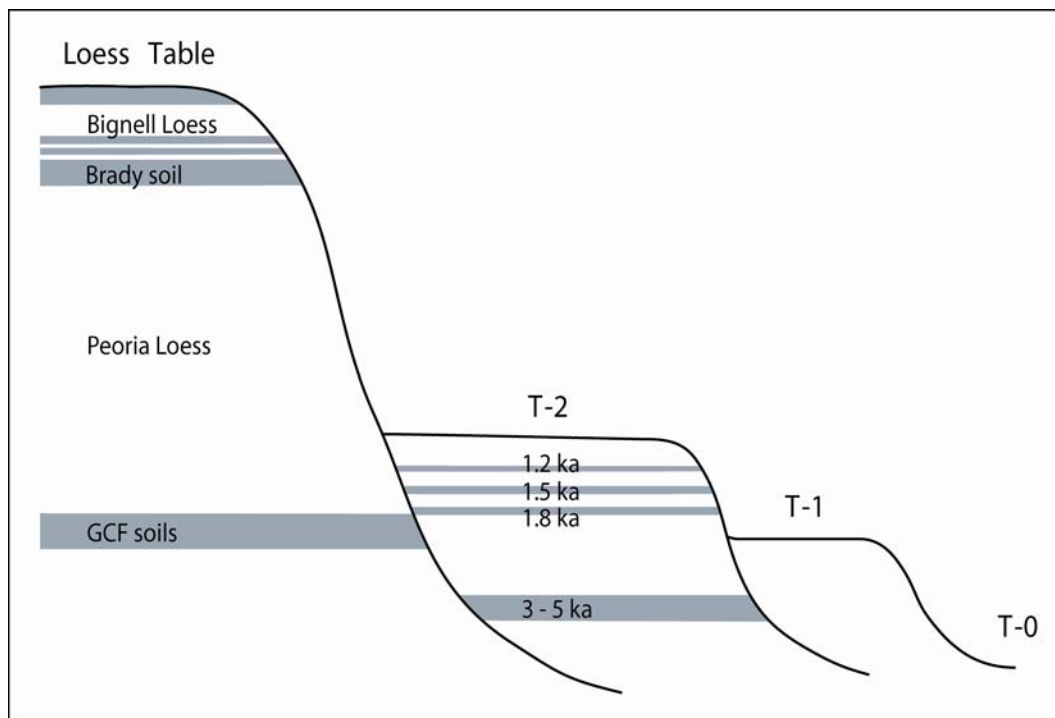


Figure 28. Generalized profiles of sediment packages in the Hay Canyon system.

Other canyons

Carol Canyon, a small canyon three miles west of Hay Canyon in the Breaks, exhibits the linear drainage pattern and more gentle topography common to canyons of the South Fork region. At the Workman site, near the upper reaches of Carol Canyon, the gully has exposed the T-2 alluvium, including a prominent buried soil dating to $1,500 \pm 80$ cal yr BP. The total depth of T-2 fill is unknown, as is the presence or absence of the older soils found at the base of the Williams and Cave sites. A test core into the T-1 fill below the Workman face yielded 1.8 m of silt with no sign of pedogenesis. Further down the canyon 0.2 km the depth of T-1 sediment was 3.1 m.

Ages derived from shallowly-buried T-2 soils of three South Fork canyons showed general agreement with the Hay Canyon model. Arnold Harvey Canyon had an Ab1 at 1.06 m dating $1,580 \pm 80$ cal yr BP, while the Ab1 in adjacent Dave Harvey Canyon dated $1,170 \pm 90$ cal yr BP at a similar depth. More detailed investigation at Hells Canyon yielded a welded soil complex dating $1,100 \pm 90$ cal yr BP for the Ab1 (0.45 m) and $1,800 \pm 80$ cal yr BP for the Ab2 (0.81 m). An Ab3 in the Hells Canyon T-2 lay under an additional 6 m of alluvium and yielded an age of $3,410 \pm 70$ cal yr BP. Stratigraphy in the South Fork canyons and Carol Canyon of the Breaks appears to support the model developed in Hay Canyon with the onset of T-2 alluviation prior to 3.5 ka, periodic stability between 1.8 and 1.2 ka, and the abandonment of that surface following 1.1 ka.

Regional alluvial records

Various researchers have reputed the scarcity of alluvial soils dating between 7 and 5 ka in the Great Plains (Johnson and Logan, 1990; Mandel, 1992a). May (1986) and Ahlbrandt *et al.* (1983) noted rapid aggradation during this period in the larger valleys of the Loup River system of Nebraska (Figure 19). Mandel (1992) reported widespread sediment removal from small watersheds across Illinois, Iowa, and Missouri associated with the Altithermal of 8 to 5 ka with concurrent alluvial storage in fans and larger valleys. Alluvial fans associated with loess canyons of the Platte River in western Nebraska record high rates of accumulation between 9 and 5.8 ka as canyons were purged of sediment (Faulkner, 2002). Knox (1983) proposed that the gentle precipitation of the Pleistocene-Holocene transition gave way to more intense cyclonic storms creating more erosive power on landscapes already experiencing major vegetation shifts. In Hay Canyon, as in regional small valleys, major erosion predated 5 ka and the development of the T-2 surface.

The Loop River in central Nebraska exhibits a wide-spread alluvial soil ranging in age from 5.7-5.1 ka (May, 2003). The Kansas River shows evidence of stabilization and soil formation around 5.0 ka (Johnson and Martin, 1987), as do the larger valleys of the Pawnee River, a tributary to the Arkansas River in western Kansas (Mandel, 1992b). A single soil at the Williams site in lower Hay Canyon was reported within this range (5,400±100 cal yr BP). Without further evidence from other T-2 stratigraphy in Hay Canyon or the South Fork canyons, it cannot be ruled

out that the soil represents the influence of the nearby Republican River, rather than that of the canyon system.

Smaller valleys of the Pawnee River in Kansas do not show stabilization until after 3.0 ka (Mandel, 1994). In Nebraska, two periods of stability on alluvial fans along the Platte River, occurring at 3.3-3.0 ka, and 1.6-1.0 ka (Faulkner, 2002) correspond with T-2 stability and the intervening aggradation noted in this study. The Republican River in Nebraska, downstream from the Breaks was investigated by Martin (1992) and revealed that following a period of incision, the Republican River experienced aggradation punctuated by stability and soil formation between 3.7 and 2.8 ka and again between 2.0 and 1.1 ka. The upper soil was covered in at least one location by coarser sediments prior to abandonment of the alluvial surface and entrenchment. The two defined periods of T-2 soil formation in the Hay Canyon model fall within the ranges reported by Martin (1992) for the Republican River, and the post-1.1 ka incision corresponds with the T-1/T-2 gully growth reported here. An investigation by Daniels and Knox (2005) in loess canyons tributary to the Republican River in Nebraska did not report a lower T-2 (labeled T-1 in their study) soil comparable to the 3.5-3.0 ka soil found at the Cave and Lower Hells Canyon sites. The timing of valley stability between 2.0 and 1.0 ka, and the post-1.0 ka entrenchment of that surface, however, fit well with the chronology found here.

Resumed stability of the Loop valley between 1.8 and 1.1 ka (May, 1992) is synchronous with this and other studies. Both aggradation and erosion is reported following 1.1 ka, with large floods and increased aridity held responsible (May,

1992). Wolf Creek, a small tributary to the Saline River in north-central Kansas also supports floodplain aggradation between 5 and 1 ka ending with punctuated stability and soil formation at *c.* 1.8, 1.5, and 1.2 ka (Arbogast and Johnson, 1994) as found in the T-2 of the Breaks and South Fork canyons. Fifteen alluvial sites in Texas and Oklahoma reported by Hall (1990) also concur with channel trenching beginning by 1.0 ka.

Regional environmental records

Northeastern Colorado dune fields, upwind from the study site, record six periods of widespread activity in the Late Holocene: 4.9, 2.4, 1.1, 0.8, 0.6-0.5, and 0.4 ka (Clarke and Rendell, 2003). The last four periods can be grouped together to characterize episodic drought between 1.1 and 0.4 ka, a time frame that fits well with the proposed period of T-1/T-2 gully growth. Dune activity around 2.4 ka falls within the period of alluviation between the lower (3.0-3.5 ka) soil and upper (1.8-1.2 ka) soil complex of the T-2. Additional sites in the northeastern Colorado dune fields investigated by Muhs *et al.* (1996) also indicate dune activation following periods of soil formation between 1.5 and 1.0 ka. Mason *et al.* (2003) reported young sand sheets in the Nebraska Sand Hills that were deposited between 1.0 and 0.7 ka. Using both sand- and loess records, Miao *et al.* (2007a) identified three periods of extensive drought in Nebraska: 9.6-6.5, 4.5-2.3, and 1.0-0.7 ka. The first falls during the Altithermal and may represent the time when the canyons were first created. The third supports evidence of severe post-1.0 ka drought reported throughout the central Great

Plains and the development of the T-0/T-1 gully. Drought between 4.5 and 2.3 ka does not readily support the model proposed here, as the T-2 alluvium was aggrading during that time and marked by soil formation at the Cave and Lower Hells Canyon sites.

The established relationships between Holocene dune activation and increased deposition of the Bignell loess formation (Mason and Kuzila, 2000; Mason *et al.*, 2003) provides additional records of environmental conditions in the region. Multiple OSL ages from the Old Wauneta Roadcut in southwest Nebraska, an unusually thick section of Bignell loess, yield a picture of periodic stability and loess flux (Miao *et al.*, 2007a). The early Holocene is marked by more than 2 m of unaltered loess, but a period of soil formation (BS3) began after 6.6 ka and ended before 3.9 ka. This range would encompass not only the 5.4 ka soil from the Williams site, but also the soil documented in the Bignell loess immediately above Hay Canyon at the Mills site (Table 4). A 0.3 m- thick soil dated between $5,670 \pm 70$ and $4,430 \pm 90$ cal yr BP, well within the age range of soil formation at the Old Wauneta Roadcut. A second Holocene soil (BS2) at that site was bounded by the OSL ages: 3,660 and 2,580 yr BP (3,605 and 2,525 cal yr BP) (Miao *et al.*, 2007a). Soils at the base of T-2 profiles at the Cave and Lower Hells Canyon sites fall within this range. A thicker, third soil (BS1) developed following 2.6 ka and spans the entire range of soil age documented in the upper T-2 of Hay Canyon and the South Fork canyons. Soil formation at Wauneta ended abruptly prior to 0.7 ka, concurrent with the abandonment of the T-2 surface. In the northern Great Plains of south-central North Dakota, a loess section

overlooking the Missouri River Valley also shows evidence of high dust flux in the early Holocene, with episodic soil development only occurring well after 5.0 ka (Mason *et al.*, 2008). After 3.0 ka, a period of upland stability at the Mirdan Canal site in central Nebraska (Mason and Kuzila, 2000), Bignell loess deposition increased in response to drought (Baker and Fredlund, 1998). This timing corresponds to the aggradation of the T-2 in this study, and may indicate an aeolian component to the fluvial sediments.

At sites around Cheyenne County, Kansas the Pleistocene-Holocene transition Brady Soil is well-expressed. At the Mills site above Hay Canyon, the Brady was dated between $11,150 \pm 90$ and $8,780 \pm 140$ cal yr BP. Closer to the Republican River, an upland section at the Jones site yielded Brady ages between $12,730 \pm 60$ and $8,310 \pm 70$ cal yr BP. A third site above the South Fork canyons, KWP, was dated between $11,470 \pm 140$ and $8,780 \pm 140$ cal yr BP. The ubiquity of a temporal equivalent to the Brady is not limited to upland landscape positions; it is also documented in a buried playa lake exposed at the Cheyenne County Landfill ($11,380 \pm 140$ cal yr BP) and in a large upland depression, Douthit Basin ($11,520 \pm 270$ and $10,760 \pm 220$ cal yr BP).

Conclusions

The Pleistocene-Holocene transition Brady Soil, which is well documented in the loess uplands (Schultz and Stout, 1945; Frye and Leonard, 1952; Dreeszen, 1970; Johnson, 1990; Johnson and Willey, 2000) with equivalent phases in other landscape

settings (Olson and Porter, 2002; Faulkner, 2002; Mandel, 2008), is not represented in the Breaks or South Fork canyons. The absence of a Brady equivalent in the canyons implies that major canyon development occurred following the upland soil genesis. Massive erosion must have taken place beginning no later than 8.5 ka and continuing throughout the Altithermal. There is no evidence yet of a T-3 surface.

The T-2 terrace surface appears to have aggraded following the Altithermal as temperatures declined to present-day values. Sediment accumulated on the stable valley floor, possibly prior to 5.0 ka but definitely before 3.5 ka and continuing until after 1.2 ka. A lower soil dating between 3.0 and 3.5 ka expresses landscape stability found throughout the region, after which resumed T-2 aggradation corresponds with dune activity in Colorado and a period of drought. A soil complex found ubiquitously within the upper 3 m of the T-2 shows stability between 2.0 and 1.0 ka, but with possible punctuations at 1.8, 1.5, and 1.2 ka. A marked hiatus of dune activity in Colorado and the Nebraska sand hills within the period lends further credence to the interpretation that ameliorated conditions dominated. Aggradation of the T-2 surface resumed, but was followed shortly (*c* 1.1 ka) by the advance of the T-1/T-2 gully.

This study agrees with the assertion by Daniels and Knox (2005) that the post 1.0 ka widespread entrenchment of large and small streams in the central Great Plains was caused by climatic forcing, and that the corresponding drought can be considered a regional expression of the Medieval Warm Period. No soils were found within the rapidly aggraded T-1; rather, stratigraphy of that terrace shows laminated sediments including lenses of Ogallala Formation and Pierre shale eroding from the headwalls

of the T-1/T-2 gullies. The T-0 sediments are presumed to be historical, as the T-0/T-1 gullies have made marked post-settlement progress.

Canyons of the Breaks and South Fork regions are Holocene features that illustrate the range of landscape response to periods of drought. Altithermal drought, with its multi-centennial duration of warmer-than-present temperatures, initiated, or at least exacerbated the canyons along the fringe of the loess plateau. The Medieval Warm Period, expressing periodic multi-decadal droughts, deepened the existing canyons. Future warming by even a few degrees or extended periods of drought caused by anthropogenic forcing could greatly intensify encroachment of canyons into the loess plateau. Likewise, small future changes in drought regime in the central Great Plains could produce exponential increases in erosion rates.

**CHAPTER 5: Carbon Isotope Variation in Modern Soils of the Tallgrass
Prairie: Analogues for the Interpretation of Isotopic Records Derived from
Paleosols**

(published as Johnson *et al.*, 2007b)

Abstract

Use of stable carbon isotope data from paleosols to reconstruct past plant community structure (C_3 vs. C_4) has become commonplace. In an effort to improve our ability to make isotope-based reconstructions and to better appreciate the pitfalls, investigations were conducted on both modern soils and paleosols in the Kansas grasslands.

Stable carbon isotope data were derived from soils and vegetation on the near-pristine C_4 -dominated grassland of the Konza Tallgrass Prairie Long-Term Ecological Research (LTER) site in northeastern Kansas. In order to evaluate the variation of $\delta^{13}C$ across the landscape, three levels of sampling were employed: 2 m-deep upland cores extracted to assess variation with depth in the soil profile, surface samples along two transects, and from 188 points within a 660 m x 690 m grid. For transect and grid points, both the upper 2 cm of sediment and the above-ground biomass were collected. Core samples taken at the Konza Prairie reveal that soil organic carbon was depleted in ^{13}C within the upper 10-20 cm relative to the remainder of the soil solum below, a phenomenon previously reported. In transects and the sample grid, soil organic carbon from soil surfaces was consistently more depleted in ^{13}C than above-

ground tissue of associated vegetation samples. Slope, azimuth, and insolation were computed from field data and a high-resolution DEM of the sample grid, but these variables offered no significant explanation of the spatial variability in $\delta^{13}\text{C}$ of soil organic carbon. The observation that modern landscape position has little effect on $\delta^{13}\text{C}$ over short distances, at least in a C_4 -dominated community, lends support to the application of $\delta^{13}\text{C}$ analysis to buried soils for paleoenvironmental reconstruction where paleolandscape position is often unknown.

Stable carbon isotope data were derived from paleosols at the Beisel-Steinle site located west of Konza in central Kansas. Three soils were examined: the interstadial paleosol within the Gilman Canyon Formation (*c* 38-27 ka), the Late Pleistocene-Holocene Brady Soil, and the modern surface soil. With the exception of isotopic depletion at the top of the surface soil, all three soils exhibited similar $\delta^{13}\text{C}$ values of about -14‰, the same values realized in the soil cores from Konza, suggesting close similarity among the plant communities of the interstade, the Late Pleistocene-Early Holocene transition, and pristine prairie of today.

Introduction

During the past 20 years, stable carbon isotope analysis has become increasingly popular among researchers reconstructing past terrestrial environments and associated climates. The earliest research in this arena was by Cerling and colleagues (Cerling, 1984; Cerling and Hay, 1986; Cerling *et al.*, 1989; Quade *et al.*, 1989; Cerling *et al.*, 1998) and also Dzurec *et al.* (1985), Schwarz *et al.* (1986), and

Krishnamurthy and Bhattacharya (1989). Due to its relatively low cost and perceived power as a proxy, stable carbon isotope analysis has since become a mainstream tool adapted to a variety of depositional environments.

Investigators working in the Great Plains of North America have embraced this technique, due in particular to a paucity of other applicable and reliable proxies for Late Quaternary climates. Regionally, stable carbon isotope analysis have most frequently been applied to buried soils, including soil carbonate (Humphrey and Ferring, 1994), carbon inclusions in opal phytoliths (Kelly *et al.*, 1991b, 1993, 1998; Fredlund, 1993), and especially soil organic carbon (Arbogast and Johnson, 1998; Boutton *et al.*, 1998; Kelly *et al.*, 1998; Johnson and Willey, 2000; Olson and Porter, 2002).

Despite the increasing number of studies employing stable carbon isotope analysis to interpret Late Quaternary vegetation and corresponding climates from buried soils, little attention has been given to isotopic relationships within modern-day analogs in the Great Plains. Studies have been conducted at a regional scale; for example Fredlund and Tieszen (1997b) provided stable carbon isotope values representing surface soil samples from 15 native prairie sites from Alberta to Texas and correlated these data with July mean temperatures. Also, Tieszen *et al.* (1997) related the C₄ grass contribution to total productivity with corresponding stable carbon isotope data using 65 sites from North Dakota to Texas. While these and other studies have documented the broad regional patterns of correlation between stable carbon isotope values in soils and those of corresponding vegetation, two other issues

need to be addressed regarding the interpretation of isotopic signals derived from modern soils. First, stable carbon isotope signals in soil profiles have been observed to change with depth (Kelly *et al.*, 1991a; Nadelhoffer and Fry, 1988; Torn *et al.*, 2002). This vertical change needs to be better accounted for in the context of sampling and interpreting isotopic data from buried soils in which a few samples are often used to represent the whole.

Second, isotopic signals of C₃ vegetation can vary according to large-scale topographic differences (Ehleringer and Cooper, 1988; Garten and Taylor, 1992; Balesdent *et al.*, 1993; Gao *et al.*, 1997), implying that the signal in the surface of the soil will vary accordingly. A key environmental factor that changes with topography is insolation, which affects microscale temperature regime and moisture availability. Galicia *et al.* (1999) found that insolation was even more important than soil physical characteristics in predicting soil moisture during the rainy season of a tropical deciduous forest. Another study conducted in China by Gao *et al.* (1997) showed that soil water and biomass production were strongly dependent on slope and aspect angles. Studies correlating carbon isotope signal variation within C₄ vegetation assemblages with topography are, however, lacking.

When sampling a buried soil for stable carbon isotope analysis, it is often difficult to ascertain the position of the sample site within the paleotopography. On occasions when it is possible to determine the site location within the buried terrain, the question arises as to whether additional paleolandscape positions need to be identified and sampled in order to reliably interpret the regional climate. High-

resolution sampling of a present-day prairie landscape has the potential to provide a sense of the degree to which stable carbon isotope values from surface soils vary according to fine-scale variations in landscape position. Consequently, additional modern analog studies in grassland environments should permit an estimation of the confidence that can be placed in interpreting paleoenvironments from stable carbon isotope data of buried soils.

To gain greater insight into the sampling of buried soils for stable carbon isotope analysis and interpreting the resulting data, particularly in the Great Plains, this study examines patterns of both vertical change within modern soil sola and variation among modern soil surfaces at different landscape positions. Research was conducted in the near-pristine tallgrass prairie environment of the Konza Tallgrass Prairie Long-Term Ecological Research (LTER) site in northeastern Kansas. Stable carbon isotope data from paleosols were obtained from the Beisel-Steinle site, a locality in central Kansas.

Stable carbon isotope theory and applications

Photosynthetic pathways and carbon fractionation

Carbon consists of two naturally-occurring stable isotopes, ^{12}C (98.89%) and ^{13}C (1.11%), which are usually expressed in ratio form ($^{13}\text{C}/^{12}\text{C}$). Isotopic fractionation occurs in nature, resulting in small variations within this ratio, the total range of which is only about 100 parts per thousand (Boutton, 1991). Plants discriminate against ^{13}C during photosynthesis such that plant tissue is isotopically

depleted in ^{13}C relative to the atmosphere (O'Leary, 1981; Boutton, 1996). Grasses in temperate climates fix carbon from atmospheric CO_2 by one of two main pathways: Calvin-Benson (C_3) and Hatch-Slack (C_4). C_4 plants discriminate against $^{13}\text{CO}_2$ less than C_3 plants (Bender, 1968; Smith and Epstein, 1971; Vogel, 1980; O'Leary, 1981; O'Leary, 1988; Ehleringer and Cerling, 2002), resulting in differential fractionation that is reflected in the carbon isotope ratio in plant tissue, and consequently in soil carbon. The ratio of $^{13}\text{C}/^{12}\text{C}$ is generally expressed as a δ value where:

$$\delta = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 10^3$$

and R equals the $^{13}\text{C}/^{12}\text{C}$ ratio of the sample or standard. C_4 plants produce a range in $\delta^{13}\text{C}$ value of about -17‰ to -10‰, with an average of -13‰, whereas C_3 plants range between -32‰ and -20‰ and average -27‰ (Ode *et al.*, 1980; Farquhar *et al.*, 1989; Cerling *et al.*, 1989; Boutton, 1991) (Figure 29). However, extreme values for C_3 and C_4 plants have yet to be determined and are responsive to environmental and genetic variation (Farquhar *et al.*, 1989). C_4 plants have a smaller range (7‰) than C_3 plants (12‰), and much of the variation within the C_4 range is explained by segregation into sub-pathways: NADP-ME (avg. -11.4‰), NAD-ME (avg. -12.7‰), and PCK (intermediate to the two) (Cerling, 1999; Kanai and Edwards, 1999). Based on anatomical and biochemical determinations, genera in the Andropogoneae tribe, such as the relatively mesic bluestems (*Andropogon* spp.), are mostly of the NADP-ME subtype, whereas genera in the Chlorideae tribe such as the xeric species buffalo

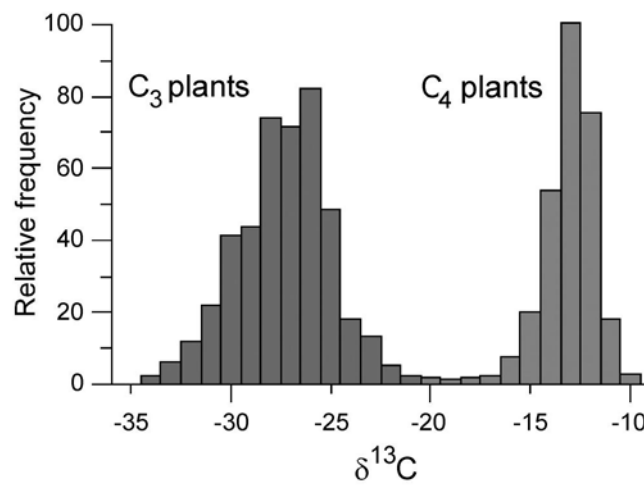


Figure 29. Relative frequency of $\delta^{13}\text{C}$ signatures of C_3 and C_4 plants. In contrast to C_4 plants, C_3 plants consistently discriminate greatly against $^{13}\text{CO}_2$ during photosynthesis, causing their tissue carbon to be depleted in ^{13}C . The average $\delta^{13}\text{C}$ of C_3 plant carbon is -27‰ , while C_4 plants average -13‰ , much less depleted. C_3 plants have a greater range in $\delta^{13}\text{C}$ expression than do C_4 plants. After Cerling (1999).

grass (*Buchloë dactyloides*) use the NAD-ME sub-pathway (Cerling, 1999). A more xeric C₄ grassland may exhibit lower (and closer to C₃) $\delta^{13}\text{C}$ values than a more mesic C₄ grassland. C₃ plants have a relatively wide range in $\delta^{13}\text{C}$, therefore, a value near the high end of the C₃ range, could conceivably be a mixture of C₃ and C₄ plant biomass, or C₃ vegetation under stress. A $\delta^{13}\text{C}$ value of about -20‰ is the lowest value assumed to indicate a C₄ environment (Cerling, 1999).

Biogeography of C₃ and C₄ grasses

Globally, the vast majority of terrestrial plants species (trees, shrubs, forbs, and cool-season grasses) employ the C₃ photosynthetic pathway, with less than 4% using C₄ (Sage *et al.*, 1999a). In spite of their small representation, C₄ plants account for about 18% of global productivity, due mainly to the dominance of C₄ monocots in grassland and steppe environments (Melillo *et al.*, 1993; Ehleringer *et al.*, 1997). Of the grasses (Poaceae), about 372 genera are C₄ and 428 are C₃, with few genera including species from both groups (Sage *et al.*, 1999a).

C₄ grasses are considered better adapted to warm, sunny, and dry environments (Barbour *et al.*, 1987; Raven *et al.*, 1992). Under conditions of full sunlight and moisture stress, C₄ plants exhibit higher light-use efficiency (LUE) than C₃ plants (Osmond *et al.*, 1982; Long, 1999). Although under ideal conditions C₃ plants are capable of a higher quantum yield for CO₂ uptake, at warmer temperatures they begin to photorespire (Ehleringer *et al.*, 1997). Increased moisture stress in C₃ plants can result in a 3-6‰ increase in $\delta^{13}\text{C}$ (Tieszen, 1991) as they lose their ability

to discriminate against ^{13}C . Additionally, early-spring burning in tallgrass prairie promotes the dominance of C_4 grasses (Gibson and Hulbert, 1987; Collins and Gibson, 1990) by removing the above-ground biomass of early-spring C_3 grasses (Harnett, 1991; Johnson and Knapp, 1995) and increasing the surface radiation receipt and temperature with the incineration of the duff layer (Knapp and Seastedt, 1986).

Since C_4 grasses rely on growing season moisture (Paruelo and Lauenroth, 1996; Ehleringer *et al.*, 1997; Collatz *et al.*, 1998), they comprise most grasses in warm to hot semiarid climates with summer rainfall maximums (Ehleringer and Monson, 1993; Amundson *et al.*, 1994). In contrast to C_3 plants, C_4 grasses have lower stomatal conductance and higher photosynthetic water-use efficiency (WUE); C_4 plants also exhibit higher photosynthetic and growth rates under high light conditions (Knapp and Medina, 1999). Despite favorable adaptations to stress, C_4 grass productivity may, in actuality, be less than that of C_3 grasses in a mixed-grass prairie (Barnes *et al.*, 1983) due to the high efficiency of C_3 photosynthesis when conditions are mesic.

Grasses and the Great Plains

While C_3 plants have a long history, grasses employing the C_4 photosynthetic pathway do not appear in the North American vegetation record until the late Miocene, as evidenced by discovery of *Tomlinsonia thomassonii* fossils in California (Tidwell and Nambudiri, 1989). C_4 grasses, including those found in the North

American Great Plains, originated in the subtropics and tropics (Paruelo and Lauenroth, 1996; Ehleringer *et al.*, 1997; Kellogg, 1999) and spread north. By 4 million years ago, C₄ ecosystems had established latitudinal and other gradients with C₃ grasses similar to those of today (Quade and Cerling, 1995; Cerling *et al.*, 1998).

High atmospheric CO₂ concentrations (above 500 ppm) favor C₃ plants (Ehleringer *et al.*, 1997; Collatz *et al.*, 1998; Cerling *et al.*, 1997), implying that the more recently evolved C₄ photosynthetic pathway is an adaptation to low levels of atmospheric CO₂ rather than moisture stress (Ehleringer *et al.*, 1991; Bowes, 1993; Cerling, 1999). Results of a plant physiology model (BIOME3) suggest that the biogeography was more affected by Last Glacial Maximum (LGM) low CO₂ levels than by LGM climate, such as in the spruce parkland of central North America south of the ice sheets (Cowling, 1999). In the event of a post-industrial rise in CO₂ concentrations though, any advantage realized by C₃ plants would be offset by increasing temperature (Ehleringer *et al.* 1997).

Regional distribution of C₄ and C₃ plants within the Great Plains has been well-documented (Terri and Stowe, 1976; Mulroy and Rundel, 1977; Stowe and Terri, 1978; Terri *et al.*, 1980), and associated vegetation assemblages mapped (Küchler, 1964, 1974). In general, the cooler climates of the northern Great Plains are dominated by C₃ grasses, such as wheatgrass (*Agropyron* spp.) and needlegrass (*Stipa* spp.). Where C₄ and C₃ grasses co-exist, the C₄ bluestems are often abundant. Tallgrass prairie of the eastern Great Plains is dominated by C₄ grasses, primarily bluestem, Indian grass (*Sorghastrum nutans*), and panicum (*Panicum* spp.). The C₄

short- and bunch grasses (buffalo grass and the gramas, *Bouteloua* spp.) dominate in the drier western and southwestern Great Plains.

Broad-scale patterns in the C₃-C₄ grass distribution are climatically forced. Several studies have demonstrated a correlation between C₃ and C₄ distribution and temperature (Terri and Stowe, 1976; Boutton *et al.*, 1980; Epstein *et al.*, 1997; Ehleringer *et al.*, 1997; Collatz *et al.*, 1998). Distribution of C₄ grassland is most closely correlated with high growing season temperature (Terri, 1988), and they exist only where mean daily minimum temperature for the warmest month is greater than about 8°C (Long, 1983). Epstein *et al.* (1997) noted that up to 81% of the variation in productivity of C₃ and C₄ grasses of the Great Plains was explained by mean annual temperature, mean annual precipitation, and soil texture. Tieszen *et al.* (1997) combined productivity of prairie remnants from State Soil Geographic Data (STATSGO), $\delta^{13}\text{C}$ values, and normalized difference vegetation index (NDVI) data to map the proportional contribution to primary production of C₄ grass biomass within the Great Plains from Texas to North Dakota. A striking feature is a pronounced zone of high C₄ contribution (>70%) about 300 km wide extending from the panhandles of Texas and Oklahoma to northeastern Kansas (Sage *et al.*, 1999b).

Plants and soil organic carbon

The $^{13}\text{C}/^{12}\text{C}$ ratio of soil organic carbon (SOC) represents an integration of the isotopic inputs from the various sources within a plant community (Troughton *et al.*, 1974; Andreux *et al.*, 1990; Balesdent and Mariotti, 1996). Further, the $^{13}\text{C}/^{12}\text{C}$ ratio

reflects the photosynthetic pathway of the dominant vegetation (Stout and Rafter, 1978; Dzurec *et al.*, 1985; Nadelhoffer and Fry, 1988) and consequently provides a reasonably good representation of SOC contributions from C₄ and C₃ plants in surface soils (Melillo *et al.* 1989; Boutton 1996; Boutton *et al.* 1998; Balesdent and Mariotti, 1996). Under circumstances in which negligible change has occurred in composition of the plant community, ¹³C/¹²C ratios of SOC in the upper several centimeters of the soil should approximate the net primary productivity of the photosynthetic pathways operating (Stout and Rafter, 1978; Nadelhoffer and Fry, 1988; Balesdent *et al.*, 1993). Mixing of carbon between litter and soil organic matter (SOM) during decomposition appears to be dynamic, in particular under C₄ prairie grasses, such as bluestem (Wedin *et al.*, 1995). Although grass roots may extend greater than 2 m (Weaver, 1968), most root biomass occurs within the upper 0.25 m of the soil solum (Jackson *et al.*, 1996), which serves to focus early stages of decomposition and integration of the plant ¹³C/¹²C ratio with that of SOC.

Should a change in the C₃/C₄ composition take place, SOM turnover rate in the soil system will determine the rate at which the SOC comes into equilibrium with the altered composition of the plant community (Cerri *et al.*, 1985; Bernoux *et al.*, 1998; Boutton *et al.*, 1998). Reservoirs of SOM in the soil have different turnover rates. SOC attached to clay- and silt-sized particles typically exhibits a slower turnover rate than that attached to sand-size particles (Christensen, 1992; Desjardins *et al.*, 1994; Bird and Pousai, 1997). A similar pattern occurs with increasing

aggregate-size variation (Jastrow *et al.*, 1996) and increasing density of particulate SOM (Golchin *et al.*, 1995).

Although researchers have documented little change in isotopic ratios of plant material during the process of decomposition and incorporation into the soil carbon reservoir (Melillo *et al.*, 1989; Wedin *et al.*, 1995; Boutton, 1996; Boutton *et al.*, 1998), microbial degradation may increase values by about 1-2‰ (O'Brien and Stout, 1978; Nadelhoffer and Fry, 1988). Several investigators have noted that SOC isotopic ratios can vary from the signal of the plant community (Rightmire and Hanshaw, 1973; Nadelhoffer and Fry 1988; Mellilo *et al.*, 1989; von Fischer and Tieszen 1995; Šantrůčková *et al.*, 2000). Overall, variations in the $^{13}\text{C}/^{12}\text{C}$ ratio smaller than 3‰ may be attributable to isotopic discrimination that can potentially occur in decomposition of SOM and with variations in carbon isotope composition of atmospheric CO_2 (Boutton, 1996). Variations exceeding 4‰ are likely attributable to a recent change in the composition of the plant community (Boutton 1996; Desjardins *et al.*, 1994).

Carbon isotope signals contained within buried soils appear to remain largely unaltered for long periods of time. Cerling *et al.* (1989) found that the 14-16‰ divergence between SOM and coexisting soil carbonate $^{13}\text{C}/^{12}\text{C}$ ratios is consistent between both modern and Miocene soils, lending credibility to the use of stable carbon isotope techniques on the record of the last 20,000 years that have been the primary focus of recent paleoenvironmental research in the central United States

(Nordt *et al.*, 1994; Arbogast and Johnson, 1998; Johnson and Willey, 2000; Nordt *et al.*, 2002).

Study areas

Konza Tallgrass Prairie LTER site

Tallgrass prairie of North America prior to European settlement comprised expansive grassland, ranging from central Kansas to Ohio and from Texas into the prairie provinces of Canada (Risser *et al.*, 1981; Samson and Knopf, 1994; Hayden, 1998). The Flint Hills (*c* 50,000 km²) contain the largest remaining tract of tallgrass prairie in North America (Figure 30) (Knapp *et al.*, 1998) and have never been plowed because, unlike the surrounding prairie region that has thick, rich soils and relatively flat topography, the Flint Hills exhibit considerable relief with mostly thin soils over bedrock containing weathering-resistant chert.

The study was conducted within the Konza Tallgrass Prairie LTER site of the northern Flint Hills (Figure 31). Watershed areas designated C4A, C1B and SB, located on the east side of the preserve, were selected based on their varied topography and different LTER-designed land management treatments. “C” watersheds are grazed by cattle, while “S” watershed is not grazed by cattle or bison. The digits “4” and “1” specify the number of years between prescribed burning of a watershed, with the last letter designation differentiating among replicate-treatment watersheds. Watersheds C1B and SB had been burned annually, while watershed C4A had last been burned in 2000 prior to spring 2003 sampling.

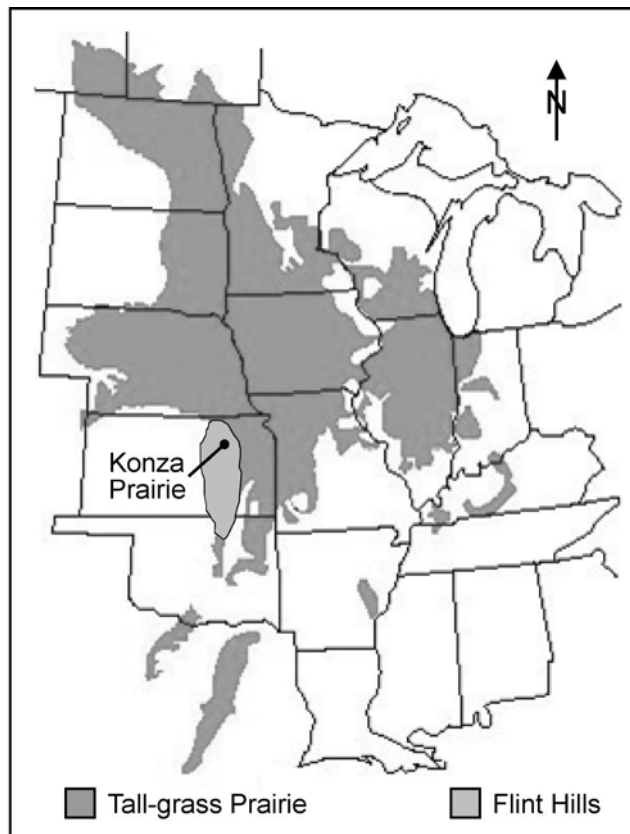


Figure 30. Tallgrass prairie in the United States and adjacent Canada. The Konza Tallgrass Prairie LTER lies within the Flint Hills, at the western edge of the tallgrass prairie in Kansas. The Beisel-Steinle site is about 180 km to the west of Konza.



Figure 31. The rolling topography of the Konza prairie, resulting from differential erosion of interbedded limestone and shale.

Geology and soils

Geologic strata at Konza are nearly flat-lying, dipping only about 0.1° to 0.2° NW (Smith, 1991). Reflected in the geology are multiple cycles of shallowing and reinundation of the Permian seas. As a result, soils on Konza have developed in marine limestone with or without chert, marine shale, or non-marine shale. In addition, alluvium, colluvium and isolated pockets of Pleistocene and reworked loess (< 2 m thick) mantle the area and provide the immediate material for soil formation.

Four groups of soils can be differentiated at the Konza Prairie based on landscape position (Wehmueller, 1996; Wehmueller *et al.*, 1994; Ransom *et al.*, 1998): (1) upland soils formed on the limestone benches, (2) soils formed on steep slopes, (3) soils on footslopes, and (4) soils on terraces and floodplains. Upland soils include fine, montmorillonitic, mesic Udertic Paleustoll; fine, montmorillonitic, mesic Typic Natrustoll; and clayey-skeletal, montmorillonitic, mesic Udertic Argiustoll (Jantz *et al.*, 1975). Soils on slopes have a wide range in depth, carbonate content, and rock-fragment content. These are loamy-skeletal, mixed, mesic Pachic Calciustoll (steep slopes); very fine, montmorillonitic, mesic Udertic Argiustoll (gentler slopes); and silty clay loams with a large amount of rock fragments (formed in colluvium). On toeslopes, soils are generally thicker (1-2 m) and are typically fine, mixed, mesic Pachic Argiustolls. On terraces and floodplains, soils are fine-loamy, mixed, mesic Cumulic Hapludolls; soils with up to 50% rock fragments; and fine, mesic Typic Calciaquolls where springs saturate the soil or fine, mixed, mesic Cumulic Hapludolls where springs do not occur (Jantz *et al.*, 1975).

Climate

Climate at Konza is temperate mid-continental with warm, wet summers and cold, dry winters. Mean annual precipitation is 83.7 cm, which is sufficient to support forest or savanna vegetation, but year-to-year variability is high (Knapp and Seastedt, 1998) and tallgrass prairie is maintained by fire and grazing, in addition to climate (Axelrod, 1985; Anderson, 1990). Even over a 10-year period, large variations occur: annual precipitation from 1991–2001 deviated between 171% and 66% from the average. About 75% of precipitation usually falls during the growing season (April through September); the wettest months on average are May and June (Hayden, 1998). Variation in average monthly precipitation is high, also, with the wettest months showing the most variability. Average annual temperature is about 13°C; the coldest month being January, with an average temperature of about -2°C, and the warmest month being July, with an average temperature of about 26.5°C. July is the month with the highest average maximum temperature (33.4°C) and the highest average minimum temperature (19.6°C). January has the lowest average maximum (4.1°C) and minimum (-7.8°C) temperatures. Gulf-derived maritime tropical air is the dominant source of precipitation and thunderstorms during April through September. Continental polar air and arctic air dominate in the winter and spring (Hayden, 1998). These air masses result in wind directions that are dominantly from the south and southeast during the late spring and summer, and from the north and west during the winter.

Vegetation

C₄ grasses dominate Konza, with the most common species being big bluestem (*Andropogon gerardii*), little bluestem (*A. scoparius*), Indian grass, and switchgrass (*Panicum virgatum*). Additionally, C₄ grasses common on mixed-grass and shortgrass prairies to the west (e.g., blue grama: *Bouteloua gracilis*; side oats grama: *B. curtipendula*; buffalo grass) occur in xeric sites on Konza. Though only a small number of the 500 species of vascular plants at Konza (Freeman 1998), C₄ grasses comprise most of the canopy cover and biomass (about 80% of aboveground net primary productivity) outside of the riparian gallery forests (Briggs and Knapp, 1995). In addition to naturally occurring C₃ forbs, shrubs, and trees, several C₃ grass species including brome (*Bromus* spp.), orchardgrass (*Dactylis glomerata*), fescue (*Festuca* spp.), and perennial ryegrass (*Lolium perenne*) have been artificially introduced at Konza.

In unburned or infrequently burned areas, woody species, especially shrubs, are prominent occupants (Freeman, 1998). These include lead plant (*Amorpha canescens*), rough-leaved dogwood (*Cornus drummondii*), fragrant sumac (*Rhus aromatica* var. *serotina*), smooth sumac (*R. glabra*), and prairie wild rose (*Rosa arkansana*). Shrubby vegetation within the study site includes rough-leaved dogwood, smooth sumac, and wild plums (*Prunus* spp.). Deciduous forest reaches up into Konza along Kings Creek, a tributary to McDowell Creek, which in turn feeds the Kansas River. Gallery forests are composed of oak (*Quercus* spp.), hackberry (*Celtis occidentalis*) and American elm (*Ulmus americana*) (van Cleve and Martin,

1991). Riparian zones within the study area occur well above any true gallery forest, and are characterized by an increase in shrubs in an otherwise open grassland.

Although no record of the vegetation history of Konza has yet been reconstructed, fossil pollen data from sites in the region suggest that during the last interglacial, grassland vegetation probably dominated the area (Fredlund and Jaumann, 1987). During the last glacial period (24 – 12 ka) spruce, mixed spruce and deciduous trees and aspen were relatively abundant, and probably existed as small stands of trees within the larger grassland. The Pleistocene-Holocene transition (after 12 ka) saw spruce abundance decline and deciduous trees and grasslands increase until about 9 ka, when grasslands once again dominated the area, as they have through the present (Fredlund and Jaumann, 1987; Fredlund, 1995; Oviatt, 1998).

Beisel-Steinle site

The Beisel-Steinle site is located on the north valley wall of the Smoky Hill River valley in central Kansas (Figure 29) about 180 km west of Konza. First reported by Johnson and Willey (2000), the site has a climate similar to that of Konza, except mean annual precipitation, at 76.0 cm, is about 8.0 cm less. Native vegetation for the area is C₄-dominated, and species composition is similar. The site consists of an 18 m-high cut along an abandoned road, which exposes a sequence consisting of a strath terrace cut in the Cretaceous Dakota Sandstone, about 2 m of locally-derived alluvial gravels resting on the bedrock, and an overlying 10 m-thick

loess mantle. Buried paleosols developed within the loess mantle at the site make it a suitable locality to examine these in conjunction with the contemporary soils and vegetation at Konza.

Methods

Field sampling

Prior to sample collection at Konza, aerial stereo photography was taken of the study area, from which a high-resolution digital elevation model (DEM) ($\leq 0.3\text{m}$ accuracy) was constructed. Using this model as a spatial reference, fine-scale variability in stable carbon isotope values was assessed in the vertical dimension through core sampling of soil profiles, and in the horizontal dimension across a landscape using transect and grid sampling.

First, eight soil cores (6 cm-diameter) were collected from areas of relatively thick loess on the upland in order to document the vertical profile of $\delta^{13}\text{C}$ values through the solum. Cores were extracted with a trailer-mounted, hydraulic soil coring machine and captured in clear plastic liners to minimize contamination and to facilitate transport to the laboratory. Next, two transects were laid out on shallow soils between topographic highs and adjacent drainage ways to determine if the sequence of near-surface limestone and shale parent materials were affecting SOM $\delta^{13}\text{C}$ values within a single aspect. Both transects had an elevation change of about 20 m. Transect “C1B” ran 198 m east-west along the south side of the boundary between watersheds C1B and C4A (east-facing aspect) and consisted of 7 points where

samples were taken of surface soil and above-ground biomass. The second transect “SB” ran southwest-northeast in the watershed (northeast aspect) and had a closer sampling interval (11 sample points within 100 m). At each transect sample-point, about 50 g of surface soil (upper 0-2 cm) were collected for $\delta^{13}\text{C}$ analysis. Corresponding vegetation samples were collected above the soil sample site by clipping, close to ground level, all plant tissue (within about 10 cm²). This method favored the collection of grasses and forbs; no tree or shrub samples were collected.

Third, a sample grid, containing points approximately 60 m apart, was applied to an area on the DEM containing portions of the three watersheds. A second grid of 60 m-spaced points was overlain on the first with a 30 m offset in both the X and Y directions. Stratified random sampling of these additional points selected 55 to be included with the original grid. The result was an approximate 660 m by 690 m area (Figure 32) with 188 sample points. The sample grid was overlain on the topography in a GIS environment to assign geospatial coordinates to each point. In the field, high-resolution GPS was used to mark each sample location for collection. Surface soil and vegetation samples were taken at each point. Vegetation samples were only analyzed, however, from a subset of 40 samples within watershed C4A. Surface soil samples and senescent vegetation were collected from the two transects and the grid before all three watersheds were burned in the spring of 2003.

At the Beisel-Steinle site, a core was extracted from the loess at the top of the exposure about 3 m from the top edge of the cut. A carbonate zone at 8.7 m halted

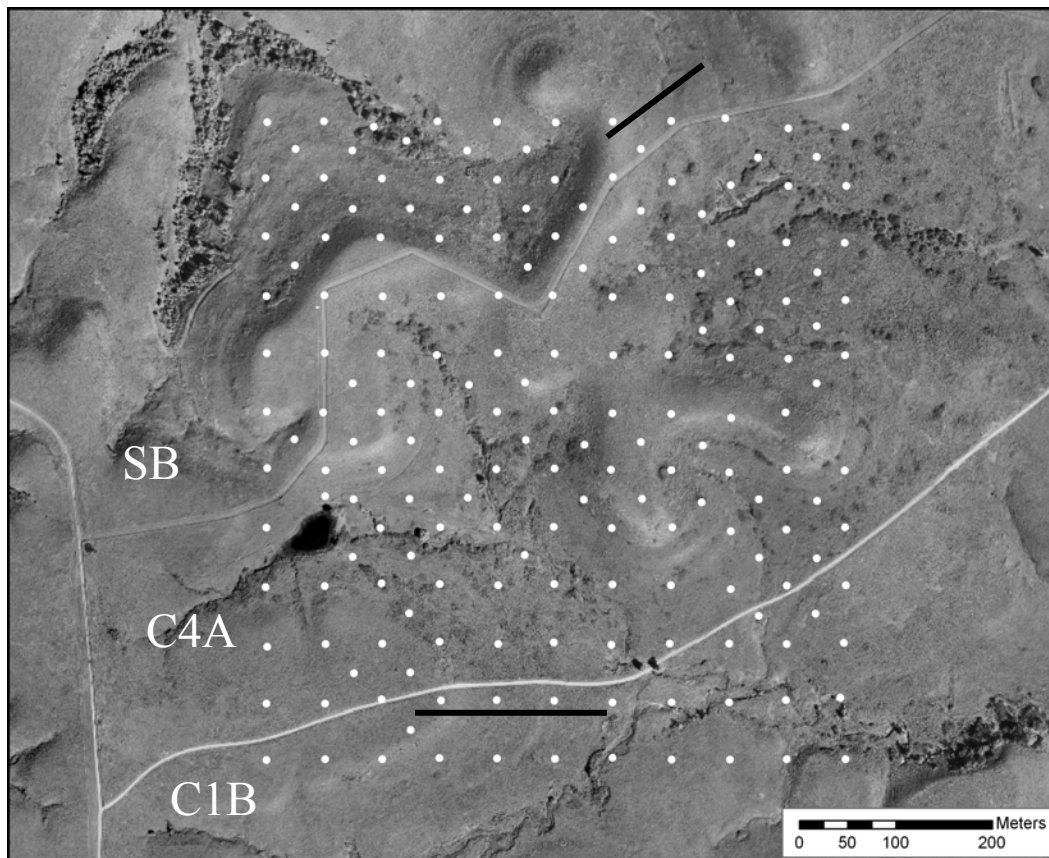


Figure 32. The landscape grid established for soil and vegetation sampling. A total of 188 sample points were selected within a 660 by 690 m area in the southwestern portion of the Konza Tallgrass Prairie LTER. Sample area is within three watersheds (divided by roads): SB, C4A, and C1B. Two sample transects are shown by black lines on the north and south edges of the sample area.

coring near the bottom of the loess mantle. As with the cores taken at Konza, these were collected and stored within plastic tubes.

Laboratory procedures

For the Konza sample grid, slope aspect values were derived for all sample points from the DEM, and, in addition, localized slope readings were taken by traditional survey methods in the field at each sample point. Slope angle, aspect, and latitude (39.1 ° N), were used to determine total incoming radiation over the course of a day. Julian day 125 (May 5), halfway between the spring equinox and the summer solstice, was chosen to characterize insolation over a growing season.

Soil cores collected in the loess mantle at Konza were described, with two selected cores being sampled every 2 cm in the upper 50 cm and every 4 cm below; these samples and surface soil samples from the transects and grid sample points were picked for rootlets, oven dried, and crushed to ≤ 1 mm. SOC was determined with a LECO CN combustion analyzer, and particle size was determined with a Beckman Coulter particle-size analyzer. The soil core collected at the Beisel-Steinle site was described and sampled for stable isotope analysis at an 8 cm interval.

In preparation for stable isotope analysis, carbonate (inorganic C) was removed from soil samples by treatment with a solution of 1 N HCl for 16 hours at room temperature while monitoring reaction and pH (Midwood and Boutton, 1997). Samples were rinsed to about pH 5.5 in de-ionized water and dried at 50°C. Soil and grass tissue samples were coarsely ground in a Wiley mill, and a sub-sample was then

ground to a fine powder in a Wig-L-Bug grinder. Powdered samples were then placed in glass scintillation vials, dried overnight at 50°C, weighted and loaded into tin capsules. Samples were combusted in a Carlo Erba 1110 elemental analyzer in the presence of helium (carrier gas) and oxygen (combustion). After drying and separating the samples by gas chromatography (GasBench II), CO₂ was passed through a ConFlo II interface and delivered with a CombiPal autosampler to a ThermoFinnigan Delta Plus mass spectrometer. Precision for all analyses was better than 0.3‰.

Samples collected for radiocarbon analysis at Beisel-Steinle were soaked in deionized water for 24 hours, with frequent agitation to break aggregates down. Samples were then wet sieved to remove all particles larger than silt size, treated with HCL as described above to remove inorganic carbon, dried at 60°C, pulverized and submitted for dating.

Results

Soil cores

Soil cores extracted from the upland loess mantle exhibit a surface $\delta^{13}\text{C}$ value of about -18‰, which is consistent with the C₄-dominated grassland environment, as exemplified by core 8 (Figure 33a). A rapid increase occurs in $\delta^{13}\text{C}$, however, to a depth of about 20 cm, where the values stabilize at -14.6‰. Values then decline with depth, stabilizing at about -24‰ within the subsoil (C horizon). This C₃-dominated signal is typical of the Late Wisconsinan Peoria Loess (Johnson and Willey, 2000) in

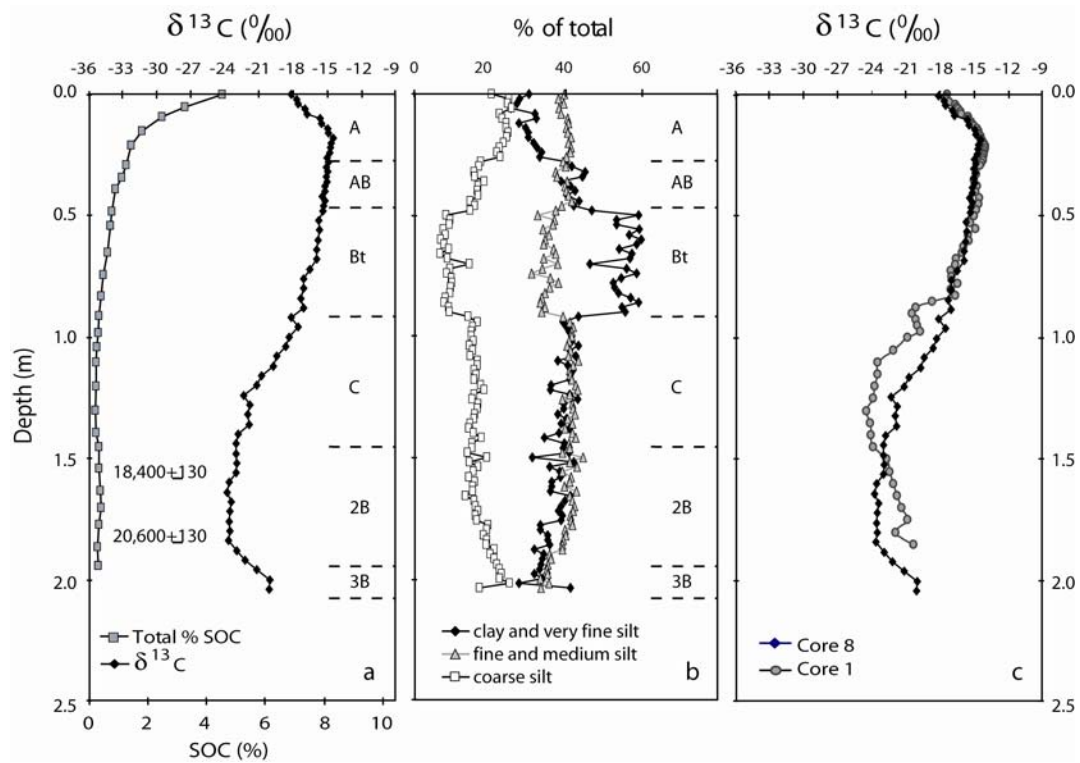


Figure 33. Vertical change in $\delta^{13}\text{C}$ and particle size distribution on representative loess uplands at Konza. (a) SOC concentration and $\delta^{13}\text{C}$ determination for core 8 collected from the loess-mantled upland. AMS radiocarbon ages ($18,400 \pm 130$ and $20,600 \pm 130$ yr BP). (b) Particle size distribution within core 8. (c) Similarity of core 1 and core 8 $\delta^{13}\text{C}$ profiles.

which the C horizon and at least some of the solum has formed. Near the base of the core, $\delta^{13}\text{C}$ values increase due to the remnants of two buried soils (2Bb and 3Bb) that most likely represent pedogenesis within later stages of the Middle Wisconsinan Gilman Canyon Formation (Reed and Dreeszen, 1965, Johnson, 1993; Johnson *et al.*, 2007a). Radiocarbon ages of $22,080 \pm 270$ cal yr BP and $24,590 \pm 290$ cal yr BP, respectively were obtained on SOC from remnants of a buried soil. SOC content decreases quickly downward in the profile from nearly 5% at the surface to less than 1% by 50 cm (Figure 33a) and appears to be independent of the $\delta^{13}\text{C}$ distribution. Particle size of the A horizon (A-AB) is dominated by fine and medium silt, whereas the best-developed part of the Bt horizon exhibits its argillic nature by a step-wise fining to about 60% clay and very fine silt (8 μm or less) (Figure 33b). At about 1.5 m and below, the 2B is revealed by a slight increase in total SOC content (Figure 33a), while the 3B development is characterized by a fining of the particle size distribution.

Core 1 produced an isotopic signature similar to that of core 8 (Figure 33c). The soil surface value of -17.3‰ corresponds to the -18.1‰ from core 8, as does the maximum value, (c 20 cm depth), -13.8‰, compared to -14.6‰ of core 8. The difference between the surface and 20 cm depth is 3.5‰ in both cores. Roots (upper 20 cm), stem, inflorescence, and leaf isotopic values from the big bluestem growing at the core 1 location yielded $\delta^{13}\text{C}$ values ranging -11.8 to -12.0‰. Above-ground bulk tissue $\delta^{13}\text{C}$ data from the same genus in Minnesota produced a comparable mean

of -11.7‰ (Wedin *et al.*, 1995). Core 1 plant tissue values are about 2‰ higher than those of the maximum soil value at a depth of about 20 cm.

Slope transects

The two topographic transects showed no discernable correlation between $\delta^{13}\text{C}$ values and topographic position (upland vs. toe slope) (Figure 34a and b). Soil samples for transect C1B ($n = 7$) ranged from -15.4‰ to -17.5‰, with a mean of -16.8‰. The average vegetation value was strongly C_4 (-13.6‰), about 3‰ higher than corresponding surface soil samples with one exception. A single sample located on a limestone bench exhibited less C_4 -dominance (-17.2‰), indicating that C_3 grasses and forbs may have gained a foothold in a microenvironment with increased moisture availability. The corresponding surface soil sample is higher than the transect average of -15.4‰. Transect SB ($n = 11$) produced similar results in that soil $\delta^{13}\text{C}$ varied closely around the mean of -16.5‰ and revealed no correlation with slope position. Vegetation samples remained about 3‰ higher than the soil samples with an average of -13.8‰. Three vegetation samples with $\delta^{13}\text{C}$ values lower than -15‰ correlate roughly with limestone benches. Unexpectedly, the vegetation sample in the toe slope position shows a very strong C_4 signal (-11.9‰) where more mesic conditions could be expected.

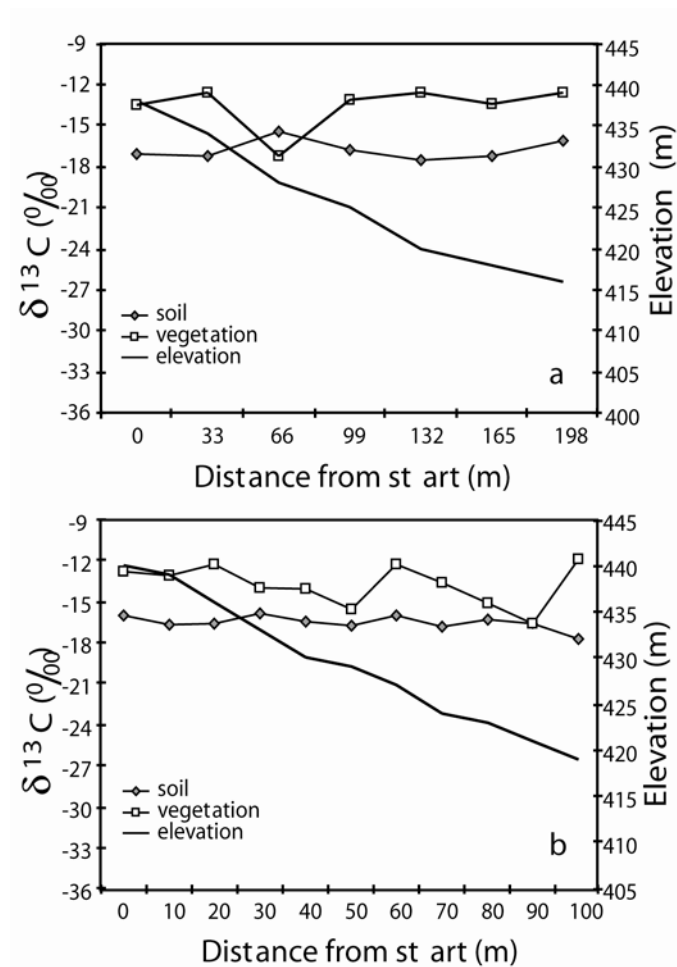


Figure 34. Soil- and grass-derived $\delta^{13}\text{C}$ values from slope transects near the landscape grid. Two transects showed no correlation between landscape position and either surface soil- or grass $\delta^{13}\text{C}$. (a) Soil samples ($n = 7$) ranged from -15.4‰ to -17.5‰ with a mean of -16.8‰ for transect C1B. Grass samples were about 3‰ higher throughout with one exception. Average vegetation $\delta^{13}\text{C}$ was -13.6‰ . (b) Transect SB ($n = 11$) produced similar results. Soil $\delta^{13}\text{C}$ varied closely around the mean of -16.5‰ and revealed no correlation with slope position. Vegetation samples remained about 3‰ higher with an average of -13.8‰ .

Landscape grid

Assays conducted on all 188 soil surface samples from within the grid resulted in a 10‰ range of SOC $\delta^{13}\text{C}$ (-14.4 to -24.3‰) with a mean of -17.5‰. Vegetation samples, analyzed from a subset of 40 sample points within watershed C4A, yielded a mean of -15.8‰, 1.7‰ higher than the grid soil values. When only the SOC samples from within C4A are considered, they produced a mean of -17.4‰, statistically similar to the mean of the entire data set. Although the vegetation samples show a characteristically pure C_4 signal, the surface soil is recording $\delta^{13}\text{C}$ values suggestive of a greater role of C_3 plants (Figure 35a).

A further disparity exists between the $\delta^{13}\text{C}$ means reported for the vegetation samples collected from transects in February (-13.7‰) and those sampled from within the C4A portion of the grid in April (-15.8‰) (Figure 35b). Surface soil samples from the two different burn treatments within the grid (C4A, burned every 4 years and SB and C1A, burned annually) showed a separation where the annual burn watersheds favored higher values, but the difference (0.7‰) was not of the magnitude found between the vegetation samples that differed both by burn treatment and sample time of year (2.1‰) (Figure 35c).

Insolation values, including both direct sunlight and diffuse light values, ranged from 30.9 $\text{mj}/\text{m}^2/\text{day}$ on steep (10-20°) south-facing slopes, to 24.7 $\text{mj}/\text{m}^2/\text{day}$ on steep north-facing slopes within the sample grid. Though surface soil samples were expected to exhibit a positive relationship between $\delta^{13}\text{C}$ values and insolation,

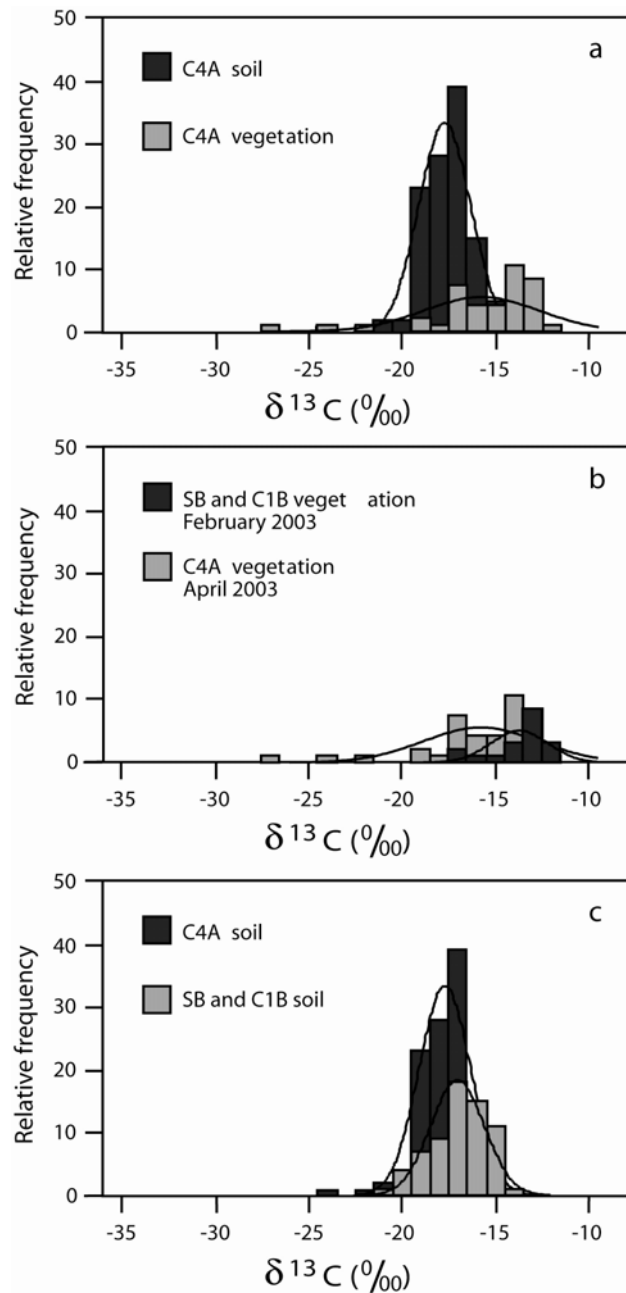


Figure 34. Normal curves and relative frequencies of $\delta^{13}\text{C}$ values. (a) Values obtained from soil ($n = 144$) and vegetation samples ($n = 40$) in the C4A watershed (b) Vegetation $\delta^{13}\text{C}$ values from the C4A watershed (collected in April 2003) are shown in comparison to vegetation values obtained on the two annually burned watersheds, SB and C1B, that were sampled from the transects in February 2003. (c) Though the difference in means is small (0.7‰), the mean value of $\delta^{13}\text{C}$ from the soils of the two annually burned watersheds (-17.0‰) is significantly higher (C.I. = 95%) than the mean from the soils in C4A (-17.7‰).

no significant relationship was revealed by regression of the two variables (Figure 36).

Beisel-Steinle

Stable carbon isotope data from the Beisel-Steinle core sample sequence and associated radiocarbon ages are provided in Figure 37. Extending up to about 6 m is the loess of the Gilman Canyon Formation. The isotopic signal of this unit averages about -21‰, a reflection of the C₃-plant environment that characterized the central Great Plains during the last glaciation. An abrupt increase in $\delta^{13}\text{C}$ above that and extending to about 3.7 m defines the paleosol formed within the loess of the Gilman Canyon Formation. This is a regionally ubiquitous paleosol which developed during Marine Isotope Stage 3 (MIS 3), an interstade within the last glaciation (Reed and Dreeszen, 1965; Johnson and Willey, 2007). Physical characteristics indicate that this paleosol is a Mollisol which was dominated by C₄ grasses (-15.7 to -13.8‰). When climatic deterioration occurred at the end of MIS 3, Late Wisconsinan Peoria Loess was deposited in a prevailing C₃ plant environment, as indicated by the isotopic values (-23.4 to -22.7‰) and by micro- and macro-fossils recovered within the region (Fredlund and Jaumann, 1987; Rousseau and Kukla, 1994). At the end of the Wisconsinan glaciation, the Brady Soil developed within the uppermost Peoria Loess (Schultz and Stout, 1945). This upper paleosol is a product of a C₄-dominated plant community and has been identified throughout the central Great Plains (Johnson and Willey, 2000). Although the Brady Soil was subsequently buried by Holocene loess,

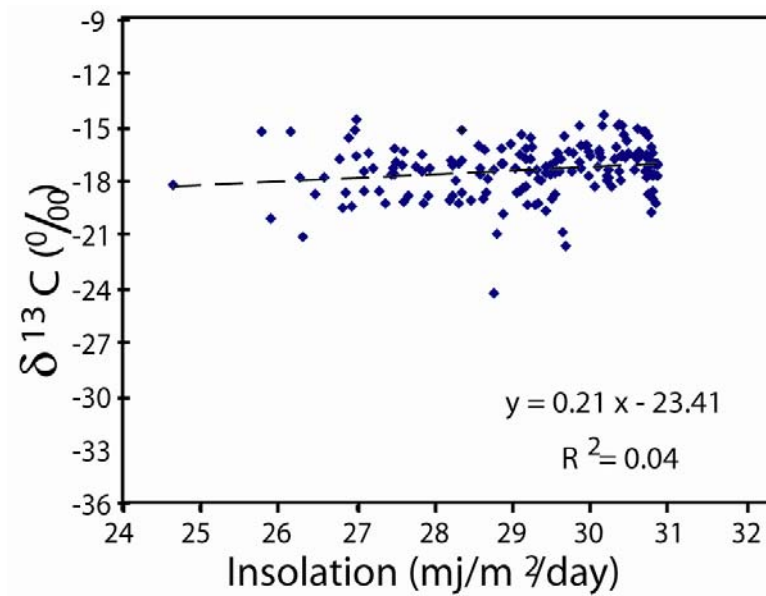


Figure 36. Effect of insolation (mj/m²/day) on surface soil $\delta^{13}\text{C}$. No significant trend was identified. The majority of samples fell well within the range typical for C₄ grasslands.

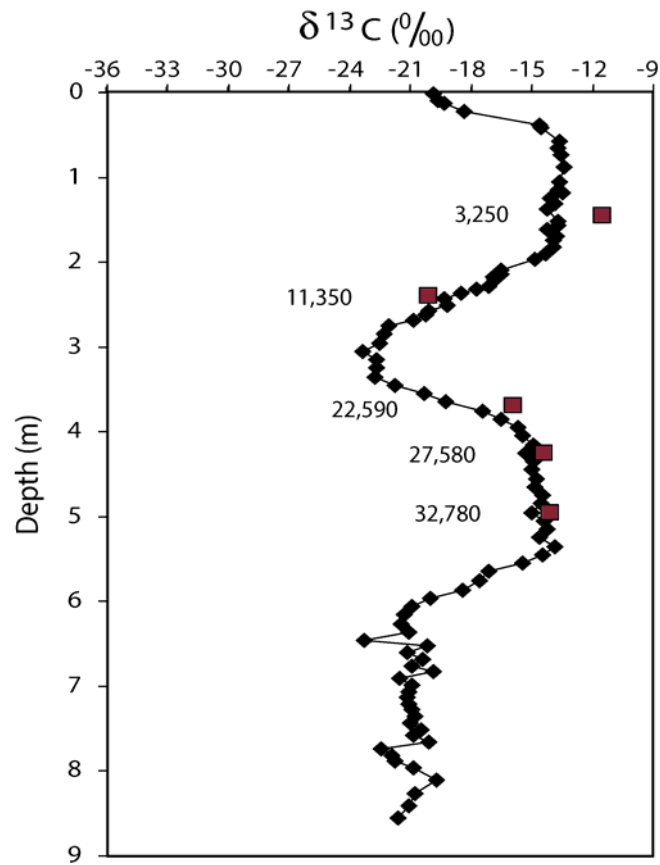


Figure 37. A loessal upland core from the Beisel-Steinle site 180 km to the west of Konza. Below 6 m the Gilman Canyon Formation loess is characterized by $\delta^{13}\text{C}$ values around -21‰, gradually increasing upward to around -15‰ as the loess transitions into the solum of the Gilman Canyon Soil. Above that, Peoria Loess values retreat to -23‰ before increasing steadily upward into the welded Brady Soil-modern surface soil complex. Gray boxes represent $\delta^{13}\text{C}$ values derived from radiocarbon ages and show good correlation with the continuous $\delta^{13}\text{C}$ curve.

the thickness at Beisel-Steinle was insufficient to separate it from the overlying modern soil, creating the interval of enriched ^{13}C values above about 2 m. The upper 40 cm of the isotope record exhibits the surface depletion recognized at Konza, but with two differences: isotopic depletion is greater at Beisel-Steinle (-19.9 vs. about -18‰) and depth of the depletion zone is thicker than as seen at Konza (40 vs. 20 cm). Both the degree and depth of depletion are greater here because the core was unavoidably taken at the edge of a cultivated field of wheat (C_3).

Five radiocarbon ages from bulk SOC obtained at the Beisel-Steinle locality record the increase in SOC age with depth (Figure 36). The ages of 38,470 cal yr BP, 32,310 cal yr BP, and 27,230 cal yr BP document the decrease in SOC age within the Gilman Canyon paleosol and indicate that it started to develop sometime prior to 38 ka, persisting until about 27 ka. The age of 13,120 cal yr BP from the lowermost Brady Soil (B horizon) dates the SOC from the earliest Brady Soil development. Above this age and within the welded A horizons of the Brady and modern soils is an age of 3,480 cal yr BP, which likely represents SOC input from modern pedogenesis. Individual sample $\delta^{13}\text{C}$ values derived by the radiocarbon laboratory to correct ages for isotopic fractionation agree relatively well with the values derived from the core sample series, the exception being the -11.9‰ obtained from the uppermost radiocarbon age (Figure 37).

Good agreement exists between modern SOC isotopic data from below the depleted interval at the two sites: a maximum of -13.4‰ here compares well with

-14.2‰ and -13.8‰ at Konza. Also, the $\delta^{13}\text{C}$ values remain relatively constant through the modern soil and the Brady Soil to which it is welded. Isotopic values at the base of the welded soil decline in a fashion similar to that of the surface soils at Konza. The -23.4‰ minimum for the Peoria Loess at Beisel-Steinle is consistent with the Peoria Loess minimum values at Konza, -24.4 and -23.7‰.

Discussion

Konza soil cores

Upward depletion of ^{13}C within the upper 20 cm and the relatively uniform enrichment downward through the B horizon documented in the soil profiles (Figure 36) suggest that factors other than the simple transference of the $\delta^{13}\text{C}$ signatures of plant tissue to SOC are involved, or that C_3 grasses are playing a yet-undetermined role. Reduced $\delta^{13}\text{C}$ values within the upper few cm of present-day soils, relative to the rest of the soil below, has been well documented (Stout *et al.*, 1981; Dzurec *et al.*, 1985; Volkoff and Cerri, 1987; Nadelhoffer and Fry, 1988; Kelly *et al.*, 1991a; Balesdent *et al.*, 1993; Bird and Pousai, 1997; de Freitas *et al.*, 2001). This observed vertical isotopic pattern in soils has been attributed to either ^{13}C depletion near the soil surface, enrichment below, or both.

^{13}C depletion

Proposed explanations for the isotopic depletion in the upper few cm of soils include a recent shift toward a more C_3 -influenced plant community, equilibration

with ^{13}C -depleted CO_2 in industrial-era atmosphere, and differential preservation and fractionation with depth. Within the upper few centimeters, SOC turns over rapidly, indicating that a shift in dominant photosynthetic pathway should be detected in a short time. For example, Martin *et al.* (1990) noted a 52-70% turnover in 16 years within a savanna soil. Wedin *et al.* (1995) observed a 1.6-2.2‰ increase in soil $\delta^{13}\text{C}$ within a C_4 monoculture after only four years, and that up to 17.8% of the soil carbon had turned over. Boutton (1996) reported a 4‰ decrease (-17 to -21‰) in SOC from the upper 5 cm of a C_4 -dominated prairie soil after 18 years of grazing (increased C_3 forbs and grasses).

Fossil fuels have a $\delta^{13}\text{C}$ value of about -27‰ since they are derived from C_3 vegetation, long before the dissemination of C_4 plants (Cerling *et al.*, 1997; Cerling, 1999). Consumption of fossil fuels has depleted atmospheric CO_2 from about -6.4‰ to -8.0‰ since the beginning of the industrial era about 200 years ago (Friedle *et al.*, 1986; Marino and McElroy, 1991; Ågren *et al.*, 1996). To place this change in perspective, about 20 ka (LGM), $\delta^{13}\text{C}$ of atmospheric CO_2 was between -7.1‰ and -6.8‰ (Lauenberger *et al.*, 1992; Marino *et al.*, 1992); $\delta^{13}\text{C}$ of atmospheric CO_2 has changed more in the last 200 years than during the last glacial-interglacial cycle. The $\delta^{13}\text{C}$ signature of atmospheric CO_2 shifts the isotopic signature of both C_3 and C_4 plants (Farquhar, 1983; Farquhar *et al.*, 1989), which should, in turn, be reflected in SOC (Balesdent *et al.*, 1993).

Tracking $\Delta^{14}\text{C}$ within a soil profile can be used to document SOC movement and turnover in a soil profile (O'Brien and Stout, 1978; Voroney *et al.*, 1991).

Boutton *et al.* (1998) noted that, for upland grassland soils in Texas, the upper 15 cm was dominated by the ^{14}C bomb spike. In comparison between the $\delta^{13}\text{C}$ profile of a 100-year old soil monolith collected from the tallgrass prairie of Russia with that of the modern soils, Torn *et al.* (2002) reported that the upper 10 cm (A1) of the modern soils exhibited an increase in $\Delta^{14}\text{C}$, indicating an imprint of the atmospheric CO_2 reservoir on the SOC in the upper part of the soil, but, since the study area is a C_3 grassland, no near-surface $\delta^{13}\text{C}$ decrease was observable in the modern soil.

SOC-derived $\delta^{13}\text{C}$ values may not reflect the isotopic plant signal due to differential preservation of biochemical fractions (Benner *et al.*, 1987). Although various components of plant tissue decompose fairly quickly, they do so at different rates (Minderman, 1968). In addition, disparate biochemical components represent a wide range of $\delta^{13}\text{C}$ values relative to that of whole-plant tissues (Dienes, 1980; Benner *et al.*, 1987). Although lignin, for example, is depleted in ^{13}C relative to the bulk plant tissue and its percentage increases as decomposition proceeds, it does not seem to deplete SOC isotopically in the upper soil (Melillo *et al.*, 1989; Wedin *et al.*, 1995).

^{13}C enrichment

Proposed explanations for the observed downward increase in $\delta^{13}\text{C}$ values include fractionation during decomposition by organisms, illuviation of ^{13}C -enriched organic matter dissolved in the soil solution, concentration of root biomass from the deeper-rooted C_4 grasses in a mixed C_3 - C_4 environment; association of SOC with the

finer soil fraction; and importation of organic carbon on aeolian particles. Several investigators have observed that organisms responsible for decomposition within the soil are more enriched in ^{13}C than the carbon they ingest (DeNiro and Epstein, 1978; Stout *et al.*, 1981; Nadelhoffer and Fry, 1988; Balesdent *et al.*, 1993). Accordingly, decomposer activity should increase $\delta^{13}\text{C}$ of SOC, and continual reprocessing of the older, deeper SOC by macro- and microorganisms may account for the enrichment of ^{13}C (Balesdent *et al.*, 1993; Boutton, 1996). In an investigation of the isotopic effect of microbial degradation in 21 grassland soils of Australia, Šantrůčková *et al.* (2000) analyzed the $\delta^{13}\text{C}$ content of SOC, soil microbial biomass, and CO_2 released by aerobic microbial respiration. They attributed the rise in $\delta^{13}\text{C}$ values with depth to be at least partially due to the stability of microbial-affected SOC and found that the isotopic effect of microbial degradation within the upper 2 cm was negated by the $\delta^{13}\text{C}$ of respired CO_2 .

Downward migration of dissolved SOM, with a relatively high $\delta^{13}\text{C}$ signature, and *in situ* enrichment from C_4 grasses have been proposed to play a role in the downward soil increase in $\delta^{13}\text{C}$. Nadelhoffer and Fry (1988), however, found no evidence for movement of organic matter using $\delta^{15}\text{N}$ as a proxy for $\delta^{13}\text{C}$. Isotopic increase with depth was attributed to *in situ* enrichment from deep-rooted C_4 grasses by Kelly *et al.* (1991a) in soil profiles of east-central Nebraska, northwestern Kansas, and eastern Colorado. Ode *et al.* (1980) noted that, in mixed-grass prairie, growth activity of C_3 and C_4 grasses are seasonally displaced, *i.e.*, C_3 plants have the highest production efficiency in late May and early June when near surface moisture is available,

whereas the C₄ plants are at their peak efficiency in late June and early July when soil moisture is at greater depth. This shift in seasonality seems to correlate with root depth (Cerling and Wang, 1996), in that the C₃ grasses tend to have a more shallow rooting system than C₄ grasses (Weaver, 1926, 1968).

SOC attached to clay and fine silt particles appears to be of greater age than that of coarser particles (Anderson and Paul, 1984; Balesdent *et al.*, 1987; Martin *et al.*, 1990; Desjardins *et al.*, 1994; Boutton *et al.*, 1998) and of lesser age with larger aggregates (Buyanovsky *et al.*, 1994; Jastrow *et al.*, 1996). As soil texture fines within the B horizon, resident SOC should be of increasing age and thereby exposed to the decomposers for a larger period of time.

For soil profiles developed in loess, such as those investigated by Kelly *et al.* (1991a), importation of organic carbon as part of the particulate load is a possibility, which would impart a baseline $\delta^{13}\text{C}$ signal to the parent material. Harden *et al.* (2002) compared ^{10}Be and ^{14}C measurements within loess-derived, grassland soil profile in southwestern Iowa and found that organic carbon in the loess originated during soil formation. Although organic compounds appear to be attached to aeolian dust prior to deposition and soil formation (Wang *et al.*, 2003), any imported organic carbon likely would be depleted, rather than enriched, in ^{13}C because of the low isotopic signal imparted to loess by the C₃ boreal environment to which it was exposed (Johnson and Willey, 2000).

Depletion and enrichment in Konza soils

Wedin *et al.* (1995) concluded from their isotopic analyses of shallow soil samples (0-25 cm) that SOM under C₄ prairie grasses is very dynamic. This observation, combined with other turnover and $\Delta^{14}\text{C}$ studies, indicates that, if the isotopic depletion in the upper soil profiles at Konza had been caused by grass cover becoming increasingly C₃ in composition, then such a change would have occurred within recent years to bring about near-surface isotopic depletion. Shift in composition did not likely occur, however, because the regime of frequent prescribed fire (1-4 year interval) maintains C₄ grass dominance.

Industrial-era ^{13}C depletion of atmospheric $^{13}\text{CO}_2$ has probably had a sufficient affect on the $\delta^{13}\text{C}$ content of the plant tissue to be transferred to the SOC, but this effect alone can not account for the 3-4‰ depletion in the upper A horizon, given that the atmosphere-induced depletion is only 1.5‰ or less (Boutton, 1996). Likewise, ^{13}C depletion from differential decomposition of lignin and other biochemical components of plant tissue does not seem to be a factor, given the research to date (Melillo *et al.*, 1989; Wedin *et al.*, 1995).

Though enrichment in ^{13}C below the upper few centimeters may be due, at least in part, to the effect of decomposers, it more likely appears to represent the dominance of C₄ grasses in this tallgrass prairie environment. The SOC isotopic signal is a result of the integration of plant-imparted $\delta^{13}\text{C}$ content over the duration of soil formation. Radiocarbon dating of soils, both modern (Boutton *et al.*, 1998; de Freitas *et al.*, 2001; Leavitt *et al.* 2007) and buried (Martin and Johnson, 1995; Muhs

et al., 1999a), documents increasing SOC age with depth. Some vertical shift may occur within the course of pedogenesis as finer soil particles concentrate in the B horizon. Below the solum, depleted isotope values reflect the environment of origin for the loess parent material.

Given the above, the most likely factor accounting for most of the near-surface ^{13}C depletion is stratification of C_3 and C_4 grass rooting systems. Though not recognized at the core sites, C_3 plants may have been present, but not apparent due to the time of year when cores were extracted. Spring growth of the shallow-rooted C_3 grasses, especially pre-burn, would be sufficient to shift the isotopic values of the upper 20 cm of the A horizon.

Konza transects and grid

Plant $\delta^{13}\text{C}$ values were typically higher than those of corresponding surface soil samples in both transect and grid sampling schemes. The difference was greatest (2.9‰) among the transect data ($n = 18$) that were collected in the annually burned watersheds (SB and C1B). By comparison, the grid samples that were taken from only C4A (burned every 4 years), showed vegetation samples ($n = 40$) to have a mean 1.9‰ above that of associated soil samples ($n = 116$). Additionally, the annually burned (transect) vegetation mean was 2.1‰ higher than the C4A vegetation mean ($n = 18$ and 40, respectively). Two known variables are at play here that are not readily separable: frequency of burning and sampling time of year. Both transects were sampled in February 2003 prior to any new growth of the vegetation. The grid

sampling however, took place in April 2003 after several weeks of new growth and prior to the spring burning. As many C₃ grasses and forbs grow aggressively in cooler temperatures. It is likely that the shift noted between the transect vegetation (-13.7‰) and the grid vegetation (-15.8‰) is due to the increasing presence of C₃ plants. The common use of fire to control the growth of C₃ plants and encourage C₄ grasses supports this supposition. This could have been tested by comparing the transect vegetation $\delta^{13}\text{C}$ values from February to grid vegetation values from the annually burned watersheds, but none were analyzed for this study. Seasonal or monthly sampling of aboveground biomass and the associated surface soil should reveal a yearly average of vegetation $\delta^{13}\text{C}$ comparable to the soil $\delta^{13}\text{C}$ value. When Boutton (1996) reported slightly higher $\delta^{13}\text{C}$ values derived from C₄ grass tissue than from soil surface SOC in un-grazed tallgrass prairie of Texas, this yearly averaging may have been a factor.

Fredlund and Tieszen (1997a) report an average $\delta^{13}\text{C}$ of -14.4‰ for eight soil surface samples collected from upland surfaces at Konza (Tieszen *et al.* 1997), compared to -17.5‰ from the Konza grid. Their average is also somewhat higher than surface assays from our upland soil cores and equal to or less than the maximum values in the cores, suggesting their samples may have been taken at a greater depth than those reported here.

Spatial variation in surface soil $\delta^{13}\text{C}$ was expected to vary with insolation, as the minority C₃ species selected more mesic microclimates on the north- and east-facing slopes, but no such relationship was found. Isotopically-depleted surface soil

values (as noted in cores 1 and 8) may have masked any direct variability of SOC $\delta^{13}\text{C}$ with insolation. Repetition of the study samples for a uniform depth (*i.e.*, 20-25 cm) may provide a more meaningful, though not quite modern, analogue. A similar surface grid in a C_3 landscape may provide insight into the variability of species composition and individual stress with topography. Since C_3 plants respond to stressful conditions by closing their stomata and using all available CO_2 (including $^{13}\text{CO}_2$) within the stomatal space, the resulting $\delta^{13}\text{C}$ of fixed carbon of C_3 plants under stress can reach higher-than-normal values. In a C_3 -dominated environment, topographic variability could reflect not only microclimates supporting pockets of C_4 vegetation, but also relative environmental stress.

Paleosol interpretations

Modern soil isotopic data from Konza provide a structure within which to interpret regional paleosols. The Gilman Canyon paleosol is isotopically pronounced, departing from the encapsulating loess units by about 7‰. While it has a mean $\delta^{13}\text{C}$ of about -14.5‰, values of -13.8‰ at the base of the Ab horizon (*c* 5.4 m) and -15.7‰ at the top (*c* 4 m) indicate a slight (*c* 2‰) shift in plant community composition during the 10 kyr of soil development. A horizon thickness is about twice that of the modern soils at Konza, which reflects its cumelic nature and long period of development. Depletion at the top of the GCF paleosol probably does not represent the shallow zone of depletion recorded in the modern Konza soils. Rather it records the transition in plant pathway from the C_4 grassland to the C_3 environment of

the LGM as Peoria Loess began to accumulate. If indeed a few cm-thick depleted zone did exist at the top of the paleosol prior to burial, it was most likely lost to detection by the compressional effects of burial and at the very least by incorporation into the basal part of the Peoria Loess through bioturbation.

Because of soil welding, the isotopic signal of the Brady Soil is partially masked. The surface soil isotopic signal, at about 0.4-1.3 m, has a mean $\delta^{13}\text{C}$ value similar to the underlying Brady Soil. Physical properties of the welded soils suggested that the original surface of the Brady Soil is at about 1.6 m, immediately below the radiocarbon sample dating to 3,250 yr BP. Isotopic data from about 1.9-1.6 m are interpreted as being a relatively uncompromised record of vegetation during the bulk of Brady Soil development. The range of -14 to -13.7‰ within the Brady A horizon is essentially the same as the surface soil and the Gilman Canyon paleosol.

Another issue when interpreting the landscape record of paleosols is that of the effect of paleolandscape position on the isotopic signal. Transect and grid data from Konza failed to define a relationship between landscape position and surface soil isotopic values. Given the lack of any discernable relationship between SOC $\delta^{13}\text{C}$ values and landscape derived at Konza, topographic position of a paleosol sampled for $\delta^{13}\text{C}$ analysis is likely not a factor that will affect the data in any significant way. One exception to this would be in the soil formed under locally wet conditions, such as a toe slope near the water table or a spring.

Conclusions

Because stable carbon isotope data from paleosol SOC are commonly used to generate reconstructions of plant communities, an improved knowledge of modern analogues would help appreciate the potential and limits of this practice. The central Great Plains have undergone dramatic changes in plant communities that have been recorded within paleosol SOC. This investigation of modern stable carbon isotope signals at Konza included documenting vertical trends in upland soils and in the soil surface and vegetation across the landscape through use of transect and grid sampling. Two soil cores from the upland loess mantle produced nearly identical isotopic profiles, which included a ^{13}C depletion zone from 0 to about 20 cm and a relatively uniform signal for the bulk of the solum below. Lowest values were at the surface, about -18‰, whereas the solum below had a maximum of about -14‰. Previous studies have also documented this near-surface depleted zone, but the processes responsible are not yet clear. Transect and grid data indicated a disparity between $\delta^{13}\text{C}$ data from above-ground plant tissue and soil surface SOC, with the former being about 2‰ higher than the latter. Also, analysis of SOC and plant tissue from grid points failed to identify a relationship between landscape position and $\delta^{13}\text{C}$ values.

As for the isotopic analysis of the Late Quaternary stratigraphic sequence conducted at the Beisel-Steinle site west of Konza, the Gilman Canyon paleosol, Brady Soil and surface soil all produced maximum SOC $\delta^{13}\text{C}$ values of about -14‰, which corresponds to that from the modern soil cores at Konza. This agreement

indicates that the plant communities of the MIS 3, Pleistocene-Holocene transition, and the present were very similar and C₄-dominated.

Three major questions have emerged from this research: (1) what are the source(s) of the ¹³C-depleted zone at the top of modern soils, (2) what caused the disparity between above ground biomass and the dominant SOC signals, and (3) would a C₃ environment a different relationship between SOC $\delta^{13}\text{C}$ signals and landscape positions.

Because surface depletion of ¹³C has been documented in soils other than just those at Konza, a strategy needs to be developed to further investigate this. If indeed this is a widespread feature of prairie soils, then surface soil sampling will consistently provide very different numbers than sampling from 25 to 30 cm. An initial effort should be a research design that better documents the geographic extent of this feature by close-interval sampling of the upper 50-75 cm of several undisturbed soil profiles at Konza; this could be expanded to sample points along east-west and north-south transects in the central Great Plains. Documentation of the plant cover in early spring and summer would be necessary to realize any spring C₃ growth and to ascertain the depth of rooting. This, in combination with isotopic analyses of soil and plant tissue samples, would provide a much better appreciation of this near-surface phenomenon.

Significant correlations between landscape position (slope aspect, slope steepness and insolation) and $\delta^{13}\text{C}$ values of SOC and plant tissue were not realized. This lack of apparent correlation may be due to the relatively small population of C₃

plants within the Konza; a greater percentage of C_3 plants would likely serve to differentiate the microhabitats as defined by landscape variation. Additional research needs to be undertaken to better evaluate patterns of SOC $\delta^{13}C$ distribution across the landscape, especially as it pertains to the isotopic signal in aboveground biomass. Rather than the coarsely sampled transect and grid used herein, a series of transects along slopes with different slope values and aspects sampled with greater frequency may better address this question. Most importantly, sample points along transects should consist of cores collected sufficiently deep to capture the isotopic character below the depleted zone. Close-interval isotopic analysis of these short cores, along with plant tissue samples, would characterize the surface and subsurface patterns.

Paleoenvironmental inferences made from $\delta^{13}C$ of buried SOM can be made without significant regard to unknown paleolandscape position. Data suggest that minor variances in insolation and associated microclimate are of less importance in determining vegetation and soil $\delta^{13}C$ than regional climate. The observed surface soil isotopic depletion is presumed to be either a modern phenomenon, or a temporary signal erased by soil burial in cumelic environments. Additional research will suggest further precautions that need to be taken when applying stable carbon isotope ratio analysis to environmental reconstruction.

CHAPTER 6: CONCLUSIONS

Periods of both Gilman Canyon and Brady pedogenesis correlate with periods of high summer insolation in the northern hemisphere (Figure 38). Previous occurrences of high insolation at about 105, 85, and 58 ka may have also been favorable for soil development, but loess stratigraphy in the Great Plains has not yet been sufficiently documented for those time periods. Luminescence dating of the pre-Sangamon Beaver Creek Soil, Buzzard's Roost soils 1-3, and the Sangamon Soil at Buzzard's Roost could perhaps further test the relationship between past insolation maximums and soil formation.

Though $\delta^{18}\text{O}$ data from North Atlantic benthic sediments (*e.g.* Martinson *et al.*, 1987) suggest much cooler ocean temperatures during Gilman Canyon development (40-25 ka) than present, C_4 vegetation dominated S3 pedogenesis as it does with modern soils of the region. This suggests that, for continental settings, strong seasonality of insolation may play a greater role than is expressed in glacial ice or ocean sediments. However, high insolation alone could also encourage C_4 grass dominance by exploiting the higher efficiency of C_4 (Hatch-Slack) photosynthetic pathway to out-compete that of C_3 plants at high-light intensity. In that case, temperatures during S3 development may have been somewhat below those of the present, while still expressing similar $\delta^{13}\text{C}$ values.

With the return of glacial conditions in the Late Wisconsinan (MIS 2; Figure 37), Peoria Loess began to accumulate and then terminated with the initial warming

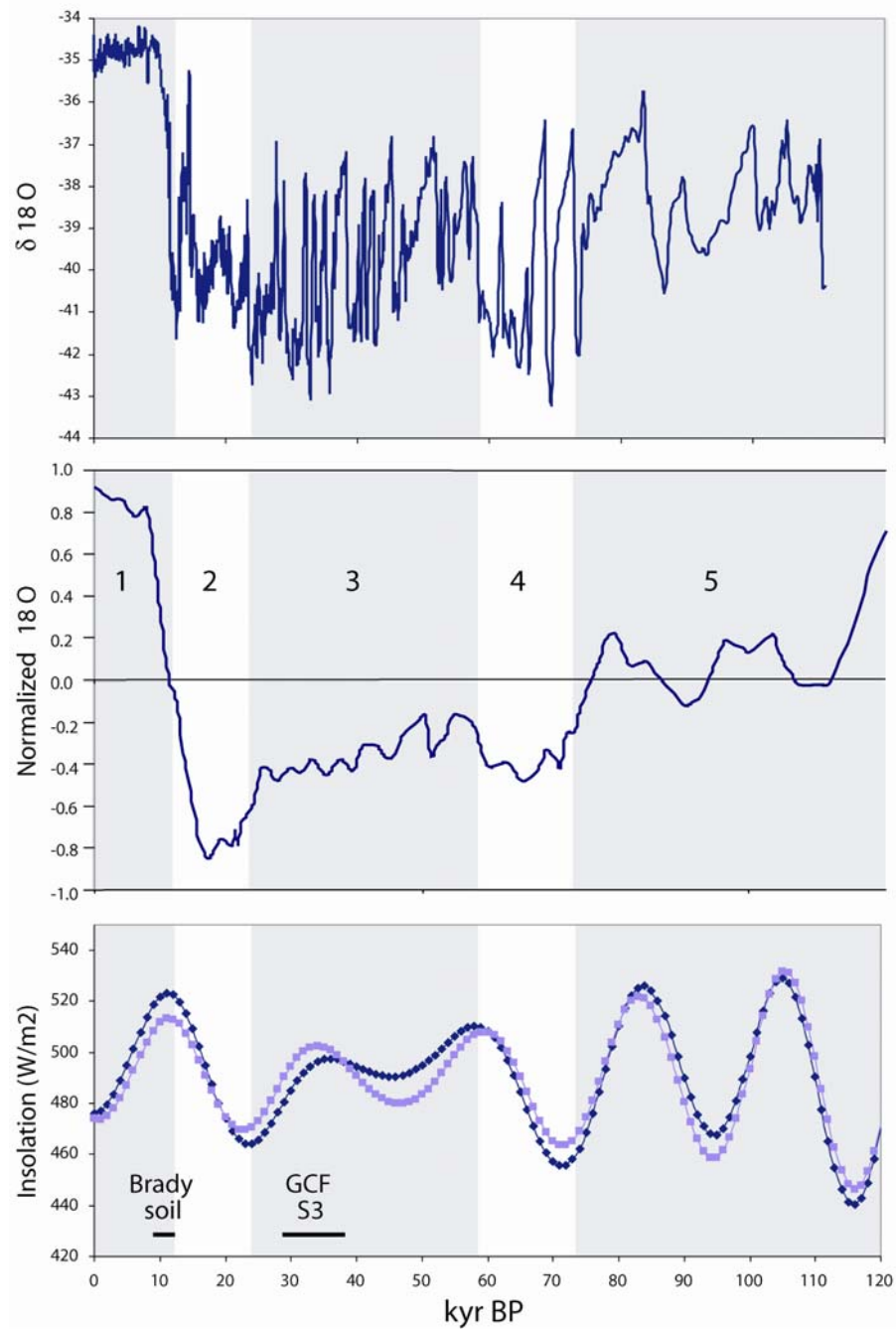


Figure 38. Temporal settings of the GCF S3 (maximum pedogenesis) and the Brady Soil. a) $\delta^{18}\text{O}$ from GISP2 (Stuiver and Grootes, 2000) b) Normalized $\delta^{18}\text{O}$ from composite deep sea cores (Martinson *et al.*, 1997); interstadies indicated by gray shading c) June insolation for 30° (purple) and 60° (blue) N latitude (Berger and Loutre, 1991).

of the current interstade. With the onset of the Younger Dryas (*c* 12.9 ka), or possibly before, dust production in the Great Plains diminished, and the Brady Soil developed over much of the landscape (Roberts *et al.*, 2003). Within the span of 2-4 kyr in the early Holocene, tree cover in the Great Plains retreated, and C₄ grasses gained a prominence that continued to the present. Throughout the north and west transects across northern Kansas, C₃ grasses were largely replaced by more the drought-adapted C₄ species. Along stream courses, riparian vegetation continued to impart a C₃ signature, (*i.e.* the DB site along the Missouri River bluffs). As with the Gilman Canyon S3, Brady Soil formation correlated closely with peak summer insolation occurring at the Pleistocene-Holocene boundary.

It has been hypothesized that the onset of the Younger Dryas was initiated by the abrupt release into the North Atlantic of glacial meltwater via the St. Lawrence Seaway (Alley and Clark, 1999). As the Gulf Stream and the global oceanic temperature conveyor consequently dissipated, global energy transfer would have slowed, allowing areas of high insolation to retain heat. Sluggish global heat transfer, in conjunction with the peak insolation for northern latitudes modeled for the Pleistocene-Holocene transition, would allow warm temperatures similar to the present in the mid-continent to contrast with a cold North Atlantic for the same period. Of note, a recent alternative hypothesis has proposed that the Younger Dryas event was related to an extraterrestrial impact (Firestone *et al.*, 2007).

Holocene warming may have had less of an impact on plant distribution in the central Great Plains than another emerging factor, *i.e.*, the Rocky Mountain

rainshadow. Loess grain size and thickness trends (Muhs and Bettis, 2003; Bettis *et al.*, 2003) document the dominant northwestern paleo-wind direction in the Pleistocene as the jet stream was diverted around the Laurentide Ice Sheet (COHMAP, 1988). With the decreasing influence of glacial ice, wind patterns may have shifted out of the south and west, causing semi-arid conditions nearer to the Rocky Mountains and ameliorating to the east.

Following the Younger Dryas, Holocene temperature variability decreased overall, but modest changes in climatic regime produced dramatic geomorphic responses in the alluvial systems of the upper Republican River. Absence of a Brady-equivalent soil in the Breaks of the upper Republican River basin suggests that the canyons may post-date Brady soil development (ending at 8.7 ka at the nearby Mills site) and are, therefore, Holocene features. Earliest ages at the base of the T-2 remnants suggest that the last phase of stability in the canyons did not begin until about 5 ka. Various proxies from the southern Plains to Minnesota and North Dakota record a period of Early Holocene drought similar to the timeframe suggested for the formation of the canyons of the Breaks (Figure 39).

Variability in initiation and termination dates between records can be attributed, at least in part, to the differing sensitivities to temperature and moisture among diverse proxies. The depression of $\delta^{18}\text{O}$ values occurring at about 8.2 ka (Figure 39) and the subsequent rise may have been a significant enough event to trigger the onset of massive erosion into the loess tablelands. Once initiated, the following warm, dry millennia may have continued to promote erosion until a

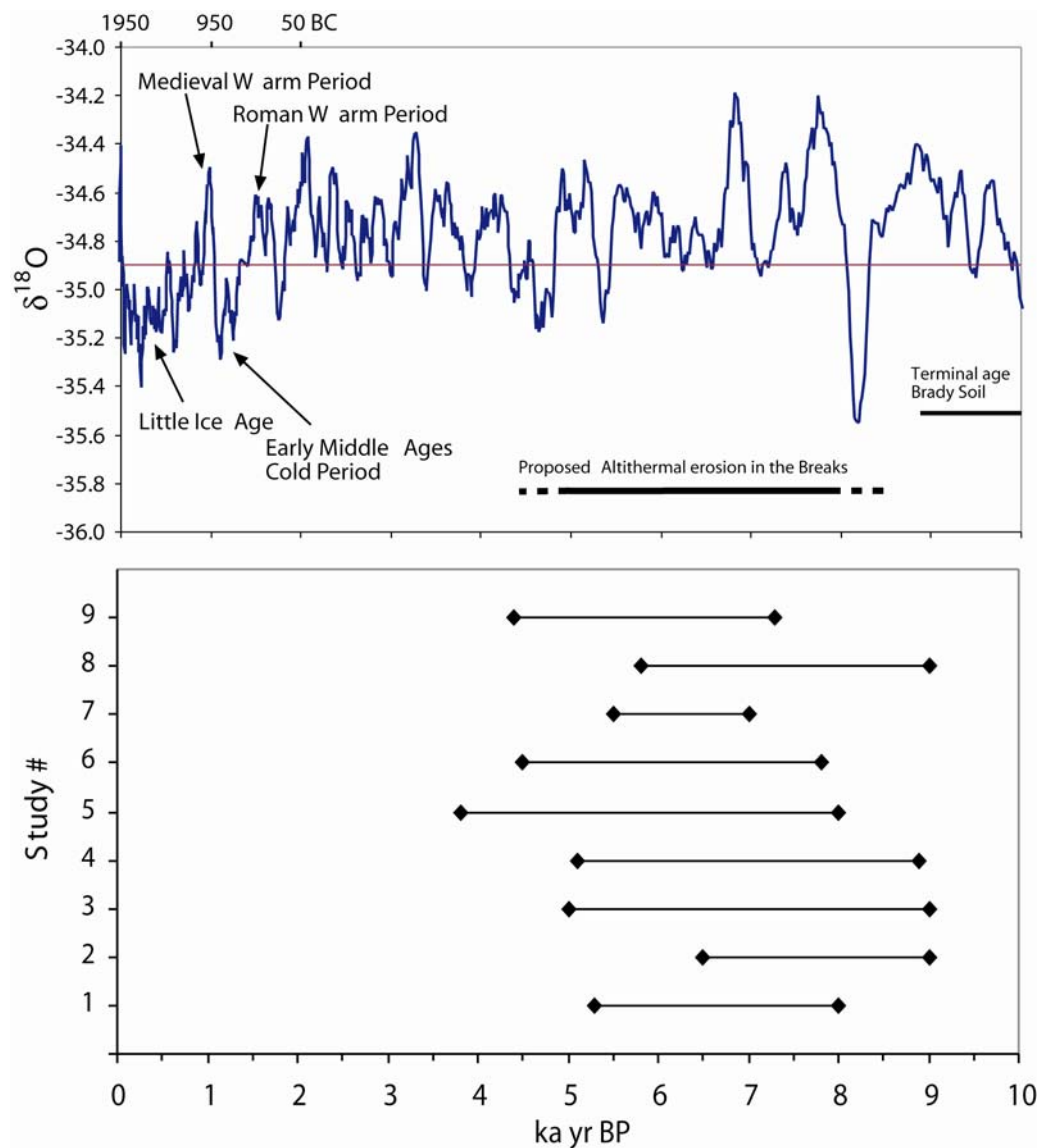


Figure 39. (a) SPECMAP $\delta^{18}\text{O}$ variability through the Holocene. (b) Regional studies indicating Altithermal warm/dry conditions. (1. Holliday, 1989; 2. Denniston *et al.*, 1999; 3. Miao *et al.*, 2005, 2007; 4. Valero *et al.*, 1997; 5. Bartlein *et al.*, 1993; 6. Dean *et al.*, 1996, 1997; 7. Forman *et al.*, 1992, 2001; 8. Faulkner, 2002; Vance *et al.*, 1992)

relatively cool spell around 4.5 ka. T-2 alluviation in the Breaks started 5-3 ka and ended abruptly across the region after 1.1 ka when gully erosion entrenched the T-2, apparently in response to the Medieval Warm Period. Modern gully (T-0/T-1) activity may have been triggered by the warming temperatures following the Little Ice age. More than just warm temperatures, abrupt warming events may be needed to trigger erosion in loess settings. Once initiated, though, degradation may continue until the next protracted period of low evaporative stress.

One further question emerges from this proposed model of Holocene climate-landscape interaction: Why did the Younger Dryas-Holocene transition yield such widespread stability (*e.g.* the Brady Soil) when the model would have predicted erosion? Possibly the high insolation and warming temperatures of that period encouraged deeply-rooted C₄ vegetation to thrive and anchor the landscape before an emerging rainshadow effect brought increasing aridity.

Stable carbon isotope relationships among modern soils, vegetation, and landscape position on the Konza LTER support the use of $\delta^{13}\text{C}$ for paleoenvironmental reconstruction. Variability with microhabitat in $\delta^{13}\text{C}$ values derived from surface soils and vegetation was negligible at Konza, lending credence to the interpretation of regional environments from paleosols using stable isotope signals derived from SOM. Repeating the study in C₃-dominated grassland may show a greater effect of evaporative stress with insolation. Additional investigation of modern soils could clarify the nature of the observed 1.6‰ difference between SOM and sampled vegetation, and likely illuminate more nuanced relationships with

climate. Future studies of the differential rooting depths of C₃ and C₄ grasses and total-year averaging of vegetation samples are both simple to design and could explain much of the divergence found at Konza between vegetation and soil profiles.

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